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CORRELATION OF COASTAL TERRACES

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ABSTRACT

The Pleistocene coastal terraces along the Atlantic seaboard of the United States are bounded by shore lines that are horizontal south of the glaciated region and that appear to continue unwarped westward along the Gulf coast to Texas. The shore lines stand at altitudes of approximately 25, 65, 95, 160, 215, and 265 feet above mean sea-level. If these horizontal shore lines owe their present height to world-wide lowerings of sea-level rather than to upheavals of the land, shore lines at the same altitudes should be recognizable along other stable coasts. Marine terraces at approximately the same altitudes have been noted in France and in South Africa. The shore lines are interpreted as records of high water due to melting of the polar ice during interglacial stages and to changes of sea-level caused by any movements of the sea bottom that may have taken place during the Pleistocene. A tentative correlation refers the 265-foot shore line to a warm pre-Nebraskan stage of the Pleistocene, each succeeding lower level to an interglacial stage, and the 25-foot shore line to an inter-Wisconsin warm period. The crustal movements which deformed Pliocene sediments but which did not warp the Pleistocene terraces mark the transition from the Pliocene to the Pleistocene.

WORLD-WIDE APPLICABILITY

A general title has been selected for this paper because the conclusions reached in it, if valid, apply not only to the broad plains along the southeastern Atlantic seaboard of the United States, with which the paper deals specifically, but to the entire world. The wave-cut or filled Pleistocene terraces of every stable coast are tentatively dated in terms of the glacial chronology of North America.

LOCAL NAMES OF TERRACES

The terraces along the coastal region of the Middle and South Atlantic States have excited much interest and speculation. As long

[1] Published by permission of the director of the U.S. Geological Survey.

ago as 1887 W J McGee[1] stated that his Columbia formation consists of a series of deltas laid down along the inland margin of the Coastal Plain by the rivers, and of a series of terraced littoral deposits connecting and graduating into the deltas and covering the remainder of the Coastal Plain to the Atlantic Ocean. Shattuck,[2] who was the first to make a detailed study of the terraces in Maryland with the aid of topographic maps, also reached the conclusion that the terraces are marine. He recognized one Pliocene terrace ("Lafayette," later called Brandywine and now referred to the Pleistocene), three Pleistocene terraces (Sunderland, Wicomico, and Talbot), and a recent terrace which is still in the process of formation. According to Shattuck, each of these terraces was formed during a period of subsidence of the land, after which it was uplifted and subjected to erosion. As Shattuck failed to recognize several of the terraces, none of the divisions that he described and mapped is a unit, and the surface of each of his terraces appears to be warped. Clark and Miller[3] accepted Shattuck's divisions and extended them into Virginia in 1912.

A series of five well-preserved terraces and remnants of two higher terraces were noted in North Carolina by B. L. Johnson[4] in 1906 and later named and mapped by Stephenson.[5] Stephenson intercalated the Coharie terrace between the "Lafayette" and the Sunderland and substituted the Chowan and the Pamlico in place of the Talbot.

Terraces in Georgia were described and mapped by Cooke[6] in 1925 and referred to shore lines stated to be approximately 60, 100, 160, 215, and 260 feet above sea-level. An additional shore line 25

[1] W J McGee, "The Columbia Formation," *Proc. Amer. Assoc. Adv. Sci.*, Vol. XXXVI (1887), pp. 221–22.

[2] G. B. Shattuck, "The Pleistocene Problem of the North Atlantic Coastal Plain," *Amer. Geol.*, Vol. XXVIII (1901), pp. 87–107; "The Pliocene and Pleistocene Deposits of Maryland," *Maryland Geol. Surv., Pliocene and Pleistocene* (1906).

[3] W. B. Clark and B. L. Miller, "The Physiography and Geology of the Coastal Plain Province of Virginia," *Virginia Geol. Surv. Bull.*, Vol. IV (1912), pp. 48–51.

[4] B. L. Johnson, "Pleistocene Terracing in the North Carolina Coastal Plain," *Science*, N.S., Vol. XXVI (1907), pp. 640–42.

[5] L. W. Stephenson, "The Coastal Plain of North Carolina," *North Carolina Geol. Surv.*, Vol. III (1912), pp. 266–90.

[6] C. W. Cooke, "Physical Geography of Georgia," *Georgia Geol. Surv. Bull.*, Vol. XLII (1925), pp. 21–35.

feet above sea-level was overlooked, because it falls on the same scarp as the next higher shore line along part of its course. The names applied to these terraces are Hazlehurst (the highest), Claxton, Okefenokee, Penholoway, and Satilla. The Okefenokee and the Satilla terraces had already been named by Veatch and Stephenson.[1]

Matson[2] described three Pleistocene terraces in Florida, the Pensacola, the Tsala Apopka, and the Newberry, with limiting upper altitudes of about 40, 60, and 100 feet, respectively, above sea-level. The altitude assigned to the Pensacola is somewhat too high, for the terrace at Pensacola is bounded by a shore line about 25 feet above sea-level. The Okefenokee and possibly higher terraces extend into the State from Georgia.

In New Jersey the names Bridgeton, Pensauken, and Cape May are sometimes used to designate terraces, but as the names were proposed for geologic formations, none of which coincides with the terrace deposits corresponding to any particular shore line, it seems best not to use these names for terraces.

HORIZONTALITY OF SHORE LINES AND THEIR ALTITUDES

Having been impressed by the remarkable horizontality of terraces in Georgia and by the close agreement of the altitudes there with those reported elsewhere, I have attempted to trace the old shore lines on topographic maps both north and south of Georgia to find out whether the terraces are really warped, as Shattuck supposed them to be in Maryland, and, if warped, how much and in what direction. I have examined maps of every State from Florida to Connecticut and a few from Mississippi, Louisiana, and Texas, but have detected little evidence of warping anywhere. If there is any warping, it is so slight as to be scarcely perceptible on maps having a 10-foot contour interval. The apparent warping in Maryland, which is in a seaward direction, appears to be due to the correlation of terraces of different ages, as explained in the foregoing, rather

[1] Otto Veatch and L. W. Stephenson, "Geology of the Coastal Plain of Georgia," *Georgia Geol. Surv. Bull.*, Vol. XXVI (1911), pp. 35–39.

[2] G. C. Matson, "Geology and Ground Waters of Florida," *U.S. Geol. Surv. Water-Supply Paper*, Vol. CCCXIX (1913), pp. 31–35.

than to actual warping of the shore line, and the discrepancies between the altitudes of terraces in neighboring States are due partly to the personal equations of the observers and partly to the fact that most of the altitudes given in the literature are those of points somewhere on the terraces and not the altitude of the actual shore line. Points on a terrace may deviate considerably in altitude from that of the shore line, being generally lower, but locally slightly higher than the mean sea-level at the time when the terraces were cut or filled.

In the area south of the terminal moraines, evidence of the presence of shore features can be detected at altitudes above mean sea-level of approximately 25, 65, 95, 160, 215, and 265 feet on almost every map whose contour interval is not too great or whose contour lines are not too much generalized. These altitudes are not precise, but are probably correct within 10 or 15 feet. It is difficult if not impossible to determine exactly the former mean sea-level corresponding to these abandoned beaches. The tidal range varied, as today, from place to place. The width and configuration of the beach, the depth of water near shore, and consequently the deviation of the foot of the scarp from the true shore line, are variable factors that have to be taken into consideration. The degree of certainty with which these factors can be evaluated depends, in large measure, upon the condition of preservation of the feature and upon the scale, the contour interval, and the accuracy of the map. A few maps of low-lying areas in Florida have a contour interval of 1 and 2 feet. There are a few excellent maps with 5-foot interval in Virginia and the Carolinas. Most of the maps examined have a 10-foot interval, but many of those from Maryland and Virginia have a contour interval of 20 feet and are very much generalized. The scale of most of the maps is 1:62,500.

Map study of terraces north of the terminal moraine is less satisfactory than south of it, for the effects of glaciation complicate the topography and obscure the terraces. Moreover, the 20-foot contour interval of maps, made when standards were less exacting than now, is too great to permit one to trace shore lines accurately without field examination. However, there are many published references to a well-developed marine 25-foot terrace along the New England coast and in the St. Lawrence Valley. In New York, terraces at altitudes

approximately the same as those in the southern States can be detected on Long Island and in the valley of the Hudson River as far north as the entrance to the Barge Canal. In the Connecticut Valley, according to Flint,[1] there is a series of terraces bounded by horizontal shore lines at altitudes of approximately 60, 100, 160, 220, and 265 feet above sea-level (as well as eleven higher terraces that range in altitude up to 860 feet) which Flint attributes to lakes impounded by temporary dams of ice. Besides these terraces, maps of the Connecticut River between Windsor Locks and Middletown and of Long Island Sound show a lower terrace which attains a maximum altitude somewhat higher than 20 feet above sea-level. The correspondence in the altitudes of these six terraces with the six in the Southeastern States is too close to be accidental. The terraces below 265 feet in the Connecticut Valley, along Hudson River, and on Long Island may have been formed at the same time and in the same manner as those south of the terminal moraines, but may have been later overridden by the Wisconsin ice sheet and veneered with glacial deposits. This suggestion is not necessarily at variance with Flint's interpretation, for lakes may have been dammed up by remnants of the Wisconsin ice to altitudes controlled by spillways over pre-existing terraces. In the Penobscot Bay area in Maine,[2] there is a widely distributed terrace from 20 to 25 feet above sea-level, a wave-built bar about 65 feet above sea-level, beach gravels up to a height of about 225 feet, but no water-washed stone above 250 feet. Antevs[3] found marine beaches from 260 to 270 feet above sea-level at several other places in Maine.

CORRELATION BY REFERENCE TO SHORE LINES

As the shore lines extend from state to state without conspicuous change of level, correlation of the local names that have been applied to the terraces becomes simply a matter of determining to which shore line each should be referred. In Florida the Newberry and

[1] R. F. Flint, "Pleistocene Terraces of the Lower Connecticut Valley," *Bull. Geol. Soc. Amer.*, Vol. XXXIX (1928), pp. 955–84.

[2] G. O. Smith, E. S. Bastin, and C. W. Brown, *U.S. Geol. Surv., Geol. Atlas, Penobscot Bay Folio*, No. 149 (1907).

[3] Ernst Antevs, "Late Quaternary Changes of Level in Maine," *Amer. Jour. Sci.*, 5th ser., Vol. XV (1928), pp. 324–27.

Tsala Apopka terraces evidently depend on the 95-foot and the 65-foot levels, respectively, and the Pensacola terrace, shown on the army engineers' map of the Pensacola quadrangle, is bounded by the 25-foot shore line. In Georgia the terrace with the 265-foot shore line has been named Hazlehurst; the 215-foot level bounds the Claxton; the 160-foot level bounds the Okefenokee; and the 95-foot level bounds the Penholoway, although the area designated as typical of the Penholoway was practically awash during the next lower stage of the sea. The Satilla terrace should be referred to the 25-foot instead of to the 65-foot shore. The unnamed 65-foot terrace in Georgia may be called by the North Carolina name of Chowan. In North Carolina the shore line corresponding to the Coharie terrace is the 215-foot level; to the Sunderland, the 160-foot; to the Wicomico, the 95-foot; to the Chowan, the 65-foot; and to the Pamlico, the 25-foot level. The so-called "Lafayette" probably includes more than one terrace.

Maryland presents more difficulties, for the names used there were not proposed for natural units. The typical Brandywine terrace appears to have a shore line at 265 feet. The name Sunderland has been applied to two terraces corresponding to the 215-foot and to the 160-foot levels, but the typical Sunderland appears to be referable to the lower level. I therefore propose to apply the name Sunderland to the terrace corresponding to the 160-foot shore line, in conformity with Stephenson's[1] usage in North Carolina, and to apply the name Coharie to the terrace corresponding to the 215-foot level in Maryland and Virginia as well as in North Carolina. The name Wicomico is usually applied in Maryland to the terrace corresponding to the 95-foot shore line, although the terrace mapped as Wicomico by Shattuck along Wicomico River appears to be the next lower. It seems best to define the Wicomico as the terrace with the 95-foot shore line, in conformity with current usage in Maryland, Virginia, and North Carolina. As Shattuck's[2] Talbot terrace includes the Chowan terrace and the Pamlico terrace, with shore lines approximately 65 and 25 feet above sea-level, the name Talbot had better be abandoned as a terrace name. In the District of Columbia and in nearby Maryland and Virginia there are at least two terraces higher

[1] *Op. cit.* [2] *Op. cit.*

than the typical Brandywine at altitudes of more than 300 and about 400 feet above sea-level, but I do not yet know how extensive they are nor whether they are warped.

GLACIAL CONTROL OF SEA-LEVEL A CAUSE OF TERRACING

Shattuck has presented convincing evidence that each period of submergence that produced a terrace was followed in Maryland by a period of emergence and erosion. This appears to have been the case in other States as well as in Maryland, for each shore line shows traces of drowned valleys. These changes in the position of the sea on the land must have been caused either by up-and-down movements of the continent or by actual world-wide oscillations of sea-level. As it seems very unlikely that such a great body of land as that over which the terraces have been traced could have lifted and sunk several times without conspicuous tilting or warping, the explanation of the phenomenon is more logically to be sought in movements of the sea, provided adequate causes for the advance and retreat of the sea can be found.

One such cause is the glacial control of sea-level. It has long been recognized that the alternate accumulation and melting of the polar ice caps during Pleistocene time must have resulted in fluctuations of sea-level because of the great volume of water piled up on the land in the form of ice during the glacial stages and restored to the sea during the interglacial stages. The glacial stages must have been times of world-wide low water; the interglacial, of high water, during each period of which the sea must have carved its high-water mark on the land. Whether or not glacial control can account for a high-water mark 265 feet above the existing sea-level is still a matter for conjecture, for we do not yet know how much water is locked up in the existing ice caps. Wright[1] has estimated that the Antarctic ice sheet alone contains enough water to raise the sea about 40 meters (130 feet) on the assumption that the average thickness of the ice is 1,000 meters. The Greenlandic ice sheet would add about 6 meters (20 feet) more.[2] Antevs,[3] assuming a greater thickness of ice, finds

[1] W. B. Wright, *The Quaternary Ice Age* (1914), p. 415.

[2] W. J. Humphreys, *Jour. Washington Acad. Sci.*, Vol. V (1915), p. 446.

[3] Ernst Antevs, "Quaternary Marine Terraces in Non-glaciated Regions and Changes of Level of Sea and Land," *Amer. Jour. Sci.*, 5th ser., Vol. XVII (1929), p. 43.

that the melting of the existing ice sheets and glaciers would raise sea-level 200 feet. If these estimates are too conservative, glacial control will account for shore lines as high as the 215-foot level.

If the fossil and recent shore lines of the Atlantic seaboard are due solely to glacial control of sea-level, the deglaciation of the polar regions must have been most complete when the highest shore line was formed. There may have been no permanent ice caps then, and every interglacial stage since must have been less icy than the present time. Such evidence as has been obtained from fossil plants and animals indicates that the climates of the interglacial stages were milder than now. However, causes other than glacial control may have contributed to a general fall of sea-level and made the shore corresponding to one interglacial stage stand lower than the shore of the next preceding, even though the ice caps may have been of the same size.

The most effective contributory causes of a fall in sea-level are the downwarping or downfaulting of part of the sea bottom. Such events may have occurred more than once during the Pleistocene. Antevs has suggested that the coastal regions may have risen isostatically because of relief from the load that has been removed from them by erosion, but so little has been eroded from the coastal part of Georgia and the Carolinas, where the terraces extend inland more than 100 miles, that the uplift due to erosion must be negligible there.

WORLD-WIDE FLUCTUATIONS OF SEA-LEVEL

A crucial test of the hypothesis that this set of coastal terraces is due to changes in level of the sea and not to movements of the land is the presence of elevated shore lines at identically the same altitudes in many widely separated regions. Evidence that such is the case is accumulating. In North America the fossil shore lines apparently have the same altitudes from at least as far north as Connecticut to Texas. Daly[1] has shown that there has been a comparatively recent world-wide lowering of sea-level amounting to 20 or 25 feet. Depéret[2]

[1] R. A. Daly, "A Recent World-wide Sinking of Ocean Level," *Geol. Mag.*, Vol. LVII (1920), pp. 246–61.

[2] Charles Depéret, "Essai de coordination chronologique général des temps quaternaires," *Comptes rendus Acad. Sci.*, Vol. CLXVI (1918), pp. 480–86, 636–41, 884–89; Vol. CLXVII (1918), pp. 418–22, 979–84.

has called attention to the fact that marine Pleistocene terraces stand at the same altitudes along the western Mediterranean, the Atlantic coast of France, and the English Channel, and has correlated them by reference to shore lines at altitudes of approximately 90–100 meters (295–328 feet), 55–60 meters (180–97 feet), 30–35 meters (98–115 feet), and 18–20 meters (59–66 feet). The presence of a gravel-covered bench 8 meters (26 feet) above mean sea-level on a chalk cliff near Calais, described by Dubois,[1] evidently indicates a shore line near that height. Perhaps all of these shore lines of France may prove to conform even more closely with those of America when the altitudes of terraces on both sides of the Atlantic shall have been more precisely determined. Along the coast of South Africa, according to Krige,[2] there are well-developed marine terraces 20 and about 60 feet above sea-level, a beach about 100 feet above sea-level at Cape Point, a shelf at 170 feet at Cape of Good Hope, one at Kogelbaai and East London at 200, and a shelf whose inner margin stands about 250 feet above sea-level in Tsitsikama. Elevated shore lines have been reported from many other parts of the world, but the altitudes assigned to them need to be critically reviewed. There are, of course, many tilted and raised terraces whose present attitudes are due primarily to local or regional movements of the land and which do not conform with the eustatic terraces under discussion.

In searching for evidence to confirm or disprove the hypothesis of world-wide changes of sea-level, one should bear in mind that the marks left by the sea during its former stands upon the land are at many places very faint and inconspicuous. The slope of the land has not been much altered from its condition before submergence, and the deposits left by the sea, except near the mouths of rivers, form merely a very thin veneer of sand or gravel which may not be distinguishable from the underlying material. The most conspicuous effects along the Atlantic seaboard are scarps which are generally only a few feet high; sandy beach ridges; old sand bars that rose not

[1] Georges Dubois, "Recherches sur les terrains quaternaires du nord de la France," *Mém. soc. geol. du nord*, Vol. VIII (1924), p. 36.

[2] A. V. Krige, "An Examination of the Tertiary and Quaternary Changes of Sea Level in South Africa, with Special Stress on the Evidence in Favour of a Recent World-wide Sinking of Ocean Level," *Ann. Univ. Stellenbosch*, Vol. V (1927), pp. 1–81.

quite to the former sea-level or stood a few feet above it as low, outlying islands; sand spits and hooks; and drainage and soil patterns suggestive of former tidal marshes.

Just as today the sea bottom is not everywhere flat or evenly sloping but has its hollows, banks, and sand bars, so the ancient sea bottoms that now stand high as coastal terraces were not featureless plains. The even slope of the modern beaches is interrupted; here, by shoals upon which the waves break far from shore; there, by deeper water which permits the waves to dash unbroken against the cliff. In general, the cliff tends to be steepest or the beach ridge highest where deep water comes close to shore, and the slope of the land is more gentle adjoining shallow water. So, in attempting to ascertain the altitude of former stands of the sea, allowance should be made for original differences in the depths of the water near shore. One cannot simply draw a shore line at the foot of every scarp, but should consider whether the sea actually reached it at mean tide or whether the water rose above its base.

COASTAL TERRACES DATED BY GLACIAL CHRONOLOGY

Students of the glacial deposits of North America recognize five principal advances of the ice sheets and call them, beginning with the oldest, the Nebraskan glacial stage, the Kansan glacial stage, the Illinoian glacial stage, the Iowan glacial stage, and the Wisconsin glacial stage. After each of these glacial invasions the ice caps dwindled in size and the oceans were flooded with water from melted ice. These warm interglacial epochs have been named the Aftonian, the Yarmouth, the Sangamon, the Peorian, and the postglacial, or Recent. Besides these five major retreats of the ice, there was at least one temporary retreat—how far we do not know—followed after a brief interval by a readvance, between middle and late Wisconsin time, which interval, for want of a better name, may be referred to as the inter-Wisconsin interglacial substage, and there was probably also a warm period at the beginning of Pleistocene time before the accumulation of the Nebraskan ice sheet.

The application of the theory of glacial control to the coastal terraces affords a possible means of correlating terraces in any stable, unwarped area with the Pleistocene deposits of the glaciated regions. If each terrace was shaped during a high-water interglacial stage, the

periods of marine low water and stream erosion would correspond to the glacial stages. If no shore lines in the series have been overlooked and if none represent merely intermediate resting stages in the retreat of the sea, it may thus be possible to trace the course of events backward, step by step, from the Recent and to assign each terrace to its proper place in the glacial chronology. On this basis one may infer that the Recent postglacial epoch was preceded by lower water during the late Wisconsin glacial substage; during the temporary retreat of the ice in mid-Wisconsin time, the water rose 25 feet above its present level; the 65-foot shore line prevailed during the Peorian interglacial stage; sea-level was low during Iowan time; the 95-foot shore line corresponds to the high water of Sangamon time; the Illinoian was a time of low water and stream erosion; the 160-foot shore line corresponds to the Yarmouth interglacial stage; the Kansan was a period of stream erosion; the sea stood about 215 feet higher than now during the Aftonian interglacial stage; and the low water of Nebraskan time was preceded by a high-water stage of about 265 feet.[1] The terraces corresponding to these stages are shown in Table I.

This tentative correlation of the Pleistocene shore lines with definite interglacial stages is highly speculative. It is based on the assumptions that the series of shore lines as now interpreted is complete; that the highest is the oldest and the lower are younger in regular sequence; that no high-water marks have been obliterated by readvances of the sea to higher levels; that none represent resting stages of the sea in its fall to lower levels due entirely to causes other than glacial control; that the current interpretation of the glacial stages in North America is correct and applies in major features to other glaciated regions. However, in partial corroboration of this speculative classification is the discovery by Leverett[2] that a gravel train derived from the Illinoian drift can be traced down the Susquehanna

[1] I suggested the correlation of the Pleistocene terraces with the interglacial stages September 19, 1928 (*Jour. Washington Acad. Sci.*, Vol. XVIII, p. 418, n.) before reading Georges Dubois' "Sur la nature des oscillations de type atlantique des lignes de rivages quaternaires," in which he attributes the American shore lines to glacial control of sea-level and correlates them with shore lines of France. Antevs (*op. cit.*, January, 1929) uses the same correlation.

[2] Frank Leverett, "Results of Glacial Investigations in Pennsylvania and New Jersey in 1926 and 1927," *Bull. Geol. Soc. Amer.*, Vol. XXXIX (1928), p. 151.

River to the head of Chesapeake Bay, where it ends in the marine Wicomico formation. According to the correlation here suggested, the deposits of the Wicomico Sea (referred to the Sangamon interglacial stage) are the oldest now above sea-level in which Illinoian débris could possibly occur. Daly has referred the 25-foot shore line to the postglacial epoch, but in New Jersey the lowest terrace appears to be overlain by a late Wisconsin terminal moraine.

ANTIQUITY OF MAN IN AMERICA

An important result of the dating of the coastal terraces is the light it throws on the antiquity of man in America. The Melbourne

TABLE I

COASTAL TERRACES OF THE SOUTHEASTERN STATES

Stage of Sea-level		Maryland	North Carolina	Georgia	Florida	Tentative Age in Glacial Chronology
Feet	Meters					
0	0.....	Recent	Recent	Recent	Recent	.....
Low water						Late Wisconsin
25	8.....	Pamlico	Pamlico	Satilla	Pensacola	Mid-Wisconsin
Low water						Early Wisconsin
65	20.....	Chowan	Chowan	Chowan	Tsala Apopka	Peorian
Low water						Iowan
95	29.....	Wicomico	Wicomico	Penholoway	Newberry	Sangamon
Low water						Illinoian
160	49.....	Sunderland	Sunderland	Okefenokee	.....	Yarmouth
Low water						Kansan
215	65.....	Coharie	Coharie	Claxton	.....	Aftonian
Low water						Nebraskan
265	81.....	Brandywine	.....	Hazlehurst	.....	Preglacial

bone bed of Florida,[1] which contains bones of man and of many extinct animals, accumulated while the sea stood at the 25-foot stage. Therefore, if the chronology here proposed is correct, man must have lived in Florida before the last advance of the Wisconsin ice sheet although some anthropologists consider him a recent immigrant to this continent.

[1] C. W. Cooke and Stuart Mossom, "Geology of Florida," *Florida Geol. Surv., Twentieth Ann. Rept.* (1929), pp. 218–26. See also *Jour. Washington Acad. Sci.*, Vol. XVIII (1928), pp. 414–21.

BOUNDARY BETWEEN THE PLIOCENE AND THE PLEISTOCENE

Although the coastal terraces show no conspicuous signs of warping, a glance at a map of the United States shows that the Atlantic seaboard has been unequally drowned. Every stream has been drowned to some extent, but those north of Cape Hatteras have been drowned deeper than those farther south. Examination of hydrographic charts brings out the fact that Chesapeake Bay and its tributaries have not been drowned as deeply as the submarine course of Hudson River. Further evidence of rather late deformation of the eastern part of the United States is given by the attitude of Pliocene deposits. Shaw[1] found considerable warping of Pliocene terraces in Mississippi; Florida has been tilted to the west since the Pliocene Citronelle formation was deposited; Campbell[2] reports that the Pliocene (?) Bryn Mawr gravel has been deformed. One is also led to inquire whether the downwarping of the Bay of Maine and of the Gulf of St. Lawrence and possibly other deformations of the glaciated regions may not have been contemporaneous with these crustal movements farther south.

The ending of the period of crustal instability which deformed Pliocene deposits and downwarped the Middle Atlantic States but which did not affect Pleistocene terraces in the same region is a convenient time at which to date the close of the Pliocene and the beginning of the Pleistocene.

[1] E. W. Shaw, "The Pliocene History of Northern and Central Mississippi," *U.S. Geol. Surv. Prof. Paper*, Vol. CVIII (1918), pp. 125–63.

[2] M. R. Campbell, "Late Geologic Deformation of the Appalachian Piedmont as Determined by River Gravels," *Proc. Nat. Acad. Sci.*, Vol. XV (1929), pp. 156–61.

ISOSTATIC COMPENSATION IN RELATION TO GEOLOGICAL PROBLEMS

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ABSTRACT

The bearing of gravity results on general geological problems is made more clear by recognizing the fact that complete local isostasy is impossible, and that such an assumption leads to appreciable errors in gravity anomalies in mountainous regions. The horizontal and vertical distributions of isostatic compensation are interrelated, and should be studied together. The ranges of assumptions as to both are stated, with the probable limits. The truth of regional instead of local isostasy has important bearings on geological deductions, and makes the gravity results more consistent with other geological evidence as to the earth's crust.

Isostasy is in large part a geological problem. Of the limited means for investigating conditions beneath the earth's surface, the most readily available are relative measurements of gravity, but for the geologist the use of this evidence is complicated by its derivation from geodetic operations, and by the elaborate computations involved. This paper will consider the results revealed by gravity measurements respecting isostatic compensation and its horizontal and vertical distribution, and the bearing of these results on the broader geophysical problems.

TEST OF THEORIES BY GRAVITY RESULTS

Dutton defined isostasy as "a condition of equilibrium of figure to which gravitation tends to reduce a planetary body."[1] This is a broad definition, and does not limit the equilibrium idea to any particular arrangement. It may still be isostasy, whether small or large unit blocks of surface area are in equilibrium, and with any conceivable arrangement of density variations with depth, so long as there is substantial equilibrium.

The evidence is abundant that there is a condition of equilibrium in the crust of the earth, and this meaning of the evidence has been understood since Airy stated it in 1855. There is no other way of

[1] Dutton, *Phil. Soc. of Washington Bulletin*, Vol. XI (1889), p. 53.

satisfactorily explaining the plumb-line deflections and the gravity anomalies.

If every element affecting the force of gravity at any point on the earth's surface were known, it would be possible to compute the theoretical force of gravity, and disregarding small observational errors, this would agree exactly with the observed force of gravity at that point. The element involving the greatest uncertainty in such computations is the distribution of the compensation, which must exist below the surface to balance the topography. Different theories as to compensation are readily tested by varying this element in the computations, and comparing the size of the resulting anomalies, or differences between observed and computed gravity. Such comparisons show which theory fits in best with the observed facts.

HORIZONTAL DISTRIBUTION OF COMPENSATION

The compensation may be assumed to be complete beneath a small unit of surface area (local isostasy), or to be distributed regionally under a larger and partially rigid surface block (regional isostasy). Vertically it may be considered as distributed beneath the surface to any depth in any conceivable arrangement. The horizontal and vertical distributions are interrelated, but for convenience the former will be first mentioned.

As the materials of the earth's crust have strength, and bear loads and resist bending stresses, it is obvious that surface inequalities will not be supported entirely locally. "The hypothesis of purely local compensation amounts to regarding the crust as made up of vertical floating piles, each able to slide freely up and down without interference from the others."[1] The impossibility of this hypothesis is not questioned, and it would not call for discussion were it not for the fact that a notable mass of gravity results has been derived on the assumption that the isostatic compensation is "complete under every separate portion of the earth's surface," however small, coupled with the conclusion that this assumption is so near the truth that any error resulting is negligible. This conclusion has been an element of mystery to geologists and others in attempting to fit together isostasy and other facts as to the earth's crust.

[1] Jeffreys, *The Earth* (2d ed., 1929), p. 199.

As I have shown elsewhere there is ample evidence from the gravity results themselves that isostatic compensation is regional, and that appreciable errors are introduced by an assumption of perfect local compensation; the original full publication of the Hayford method, in a supplementary paragraph, itself confirms these statements. The error becomes appreciable in mountainous regions, where summit stations show by too large positive anomalies the effect of overcompensation and valley stations show the reverse.

The extent of blocks regionally compensated is difficult to determine accurately from gravity results, but there is fairly good evidence that it is considerable, and may be of the order of disks of 100 miles radius. There is no reason to suppose, however, that there is any regularity or uniformity in this respect. Undoubtedly the extent of regionality in compensation for any particular region will depend on such variable factors as the strength of the crustal materials, the thickness of the crust, and the existence of faults. But there can be no question that the load of a mountain, for example, is distributed beyond the area of its base.

VERTICAL DISTRIBUTION OF COMPENSATION

Compensation is effected by material of density sufficiently lighter or heavier than the average, to balance the surface inequalities; in effect, it is a negative attraction. The compensation or density differences may be conceived to be distributed in any one of an unlimited number of arrangements beneath the surface, but the actual arrangement must lie within reasonable and rather restricted limits.

The Hayford reduction results, constituting the principal available data, are all derived on a hypothesis of a uniform distribution of density in each very small unit column, with all columns extending to a uniform moderate depth. It has not been claimed that there is any evidence that this vertical arrangement is more likely to be true than some other hypothesis. General consideration of the problem of the earth's crust indicates that even an approximation to such an arrangement of densities is quite improbable, as respects either uniform depth or uniform density vertically. The arrangement is an imaginary one, assumed for convenience, and the results are valuable for comparative study.

On this hypothesis the total uniform depth to which the compensation extends has been computed from plumb-line deflections and from gravity anomalies. All of these results, however, are based also on the assumption of complete local compensation. From the nature of the case, the evidence as to depth of compensation depends principally on results in mountainous regions. But it is just in these regions that the errors resulting from the assumption of local compensation become significant. It is therefore evident that depths of compensation thus derived will differ from depths computed on a correct hypothesis of some degree of regional isostasy. The overcompensation of summit stations will be reflected by too large a depth of compensation. This effect is shown in the varying results for the depth, derived from stations grouped according to topographic location.

A simpler assumption for study of this problem of vertical distribution would be that of an "effective depth of compensation," at which the whole compensation should be considered as concentrated; this would be a frankly imaginary conception, and as such would avoid apparent conflict with the realities of nature. This idea will be followed in this paper.

RELATION OF REGIONALITY AND DEPTH OF COMPENSATION

Regionality and depth of compensation have heretofore been treated as separate problems, but they are necessarily interrelated, and must be considered together.

Assuming some regionality of compensation instead of complete local compensation has, in general, the effect of diminishing positive gravity anomalies for stations above the average level ("above" stations), and an opposite effect for stations below the average level ("below" stations). This necessarily follows from a leveling-off process, or horizontal distribution of compensation, instead of placing all the compensation for a feature directly beneath that feature.

On an assumption of local isostasy, increase in the assumed effective depth of compensation has the effect of diminishing positive gravity anomalies (or increasing negative anomalies). This effect is greater numerically for "above" stations than for "below" stations, because the compensation correction affected is larger for the "above" stations. An assumption of greater depth with local com-

pensation has also the effect of more nearly equalizing the compensation effect between adjacent summit and valley stations (as shown in Fig. 1); this follows from the fact that the deeper the compensation is placed, the more nearly uniformly does it act on summit and on valley.

Increasing the depth of local compensation, therefore, appears to have an effect similar to that caused by increasing the regionality. While for a deep-seated local compensation there may appear to be

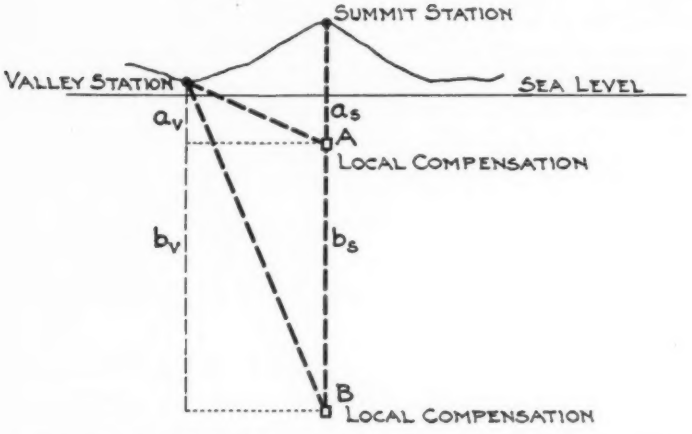

FIG. 1.—The regional effect of increasing the assumed depth of local compensation. The effect on g is the vertical component only, and becomes more uniform for the two stations, as the assumed depth is increased.

a regional compensation at less depth which is substantially equivalent, it must be noted that the former is only an imaginary arrangement. Any approach to deep-seated local compensation is impossible in the crust of the earth. With crustal materials having any strength, it is evident that compensation, if strictly local, would have to be concentrated at the surface. Local compensation departs further from the possibilities of nature the deeper it is conceived to be below the surface.

The relation of regionality and depth is further shown by the limiting cases. The minimum for each is complete local compensa-

tion, concentrated at the surface (or more strictly in a horizontal plane through the station). The maximum for depth is compensation concentrated at the center of the earth, and the maximum for regionality is uniform compensation for the entire surface; in either case this maximum corresponds to a completely rigid crust for the whole earth, so that local compensation at the earth's center becomes equivalent to regional compensation for the whole surface. In the gravity reductions these limiting cases are represented by the free-air and the Bouguer reductions, respectively. The Hayford reduction assumes complete local compensation extending to a particular depth of 70 miles. My average elevation reduction of 1895 is a regional compensation for areas of 100 miles radius and theoretically is near the truth, although only approximate as respects results for individual stations. Anomalies computed on these four different hypotheses are available as well as on some other assumptions.

For a moderate extent of regional compensation, it is possible to separate the effects of regionality and of effective depth, by the analysis of gravity results in mountainous regions, and particularly by the comparison of pairs of neighboring summit and valley stations, because of the differing effects on "above" and "below" stations of varying these two elements.

The present results for so-called "depth of compensation" are not valid, because they are based on an assumption of complete local compensation, which introduces appreciable errors as previously stated.

Hayford originally derived the depth of compensation from plumb-line deflection anomalies, assuming complete local isostasy, and uniform density vertically; he showed, however, that other assumptions as to depth and density distribution would give as good results. The average elevation gravity reduction, without any assumption as to depth or vertical distribution, gave as small average anomalies as the Hayford reduction. This illustrates the fact that as regionality is increased in gravity reductions, results for depth of compensation will become more indefinite; this follows from the principle of the attraction of an extended horizontal plate.

The problem requires that the effective depth of compensation be

computed for different degrees of regionality. It is not likely, however, that gravity results will yield definitive results as to vertical distribution, other than an approximate value for the "effective depth of compensation." Although an unlimited number of combinations of depth and regionality of compensation may be conceived of, it is probable that the actual conditions existing in the crust lie within limits which are quite moderate as compared with the two principal dimensions of the earth, that is, that the compensation lies relatively near the surface, and that the effective rigidity extends over moderate areas only.

GEOLOGICAL DEDUCTIONS FROM REGIONAL ISOSTASY

The truth as between regional and local isostasy is of fundamental significance to the geologist in studying the general condition of the earth's crust. The apparent conflict between the results of a strictly local reduction method and the generally accepted ideas of regional isostasy has led to some confusion in interpreting the evidence from gravity measurements. The earlier publications of the Hayford reduction results gave the maximum regional effect as a zone of about 12 miles radius, or an area of about 500 square miles. This is less than one-sixtieth part of the block of over 30,000 square miles, now indicated as possibly having some regional compensation; the Hayford anomalies are actually computed on the basis of completely compensated blocks or columns having a surface area so small that 200,000 such blocks equal 1 square mile.

Regional isostasy is a conception of the earth's crust quite different from complete or nearly complete local isostasy, as shown by the following differences in deductions:

Regional isostasy is in harmony with other facts as to the strength of the crustal materials. The known facts as to rock strength cannot be fitted in with complete local isostasy, which requires an assumption that small columns of rock are free to move vertically, but not horizontally.

Equilibrium requires that unit prisms of the crust have equal masses. This can evidently be true for surface blocks of considerable area, but not for unit prisms of very small cross-section. A small unit prism under the summit of a mountain will in general not be in

equilibrium with a unit of the same area beneath an adjoining valley, but the relations of these two will be maintained, to some extent at least, by the strength of the crust. As a corollary to this, the density in small unit columns will not necessarily vary so as to equalize the masses in columns having different surface elevations.

Regional isostasy fits in with reasonable conceptions as to crustal action under readjustment. Where a fracture in the crust occurs, from any cause, stress in that region, due to the tendency to equilibrium, may be relieved, and the fault line may become the limit of regionally compensated blocks. The relieving of the stress along the line of breakage of two blocks might somewhat tilt the direction of their nearly vertical movements, relatively to each other, and there would here be found a cause for a rifting apart or crushing together under isostatic readjustment or other force. The tilting which recently is being measured in portions of the earth's surface may to some extent be associated with regional isostatic readjustments. It is impracticable to apply these ideas on a theory of close local compensation, with a crust so weak that small vertical columns are in equilibrium.

Regional compensation corresponds to a crustal strength capable of bearing considerable loads, and in this respect is more consistent than is local isostasy, with visible evidence on the earth's surface.

The idea of regional isostasy has a bearing on the so-called Airy and Pratt theories. Local surface inequalities are in large part carried by the strength of the earth's crust, and the compensation is spread out more or less uniformly for considerable areas, with perhaps bulges, rather than with downward projections to support individual features; Airy referred to plateau blocks 100 miles across. The differences between Airy and Pratt are less impressive than has been thought. Airy also assumed that greater densities underlie the oceans, and such density variations would tend to equalize the crustal depths. Pratt accounted for conditions in part by horizontal contraction of the crust. Two ideas have been set up as the Pratt theory, expansion and contraction vertically as an important explanation of surface topography, and a uniform depth of compensation without projections into a denser substratum. These two ideas are independent of each other, and they should be examined sepa-

rately. In fact, it appears that vertical expansion could as well work downward as upward, and if so it also might cause projections of crustal material into a denser substratum.

NOMENCLATURE FOR GRAVITY ANOMALIES

Gravity anomalies are meaningless, or misleading, unless they are labeled to show the theory on which they are derived. Thus the statement that gravity is excessive in the Hawaiian Islands has no significance unless it is coupled with explanation of how the anomalies under consideration are derived. There are gravity stations in these islands on summits very high above the surrounding ocean bottom, from which they rise with great abruptness. The gravity anomalies for such stations are materially affected by the question of regional or local compensation, as is shown in the results below for Mauna Kea. This mountain station is 13,060 feet above sea-level, and 29,830 feet above the bottom of the sea within 60 miles. So far as the evidence from this one outstanding station is concerned, gravity in the Hawaiian Islands may be fairly normal; on the assumption of regional isostasy, the anomaly is not greater than local effects found elsewhere. The excess in gravity that has been mentioned in these islands is apparently in large part due to the erroneous assumption of complete local compensation

Reduction Method	Gravity Anomaly for Mauna Kea Observed g 978.060 Minus Computed g
Free-air reduction (isostatic, with complete local compensation at surface)	+0.628
Hayford reduction (isostatic, with complete local compensation at depth of 70 miles)	+0.183
Average elevation reduction (isostatic, with regional compensation to a radius of 100 miles, and no assumption as to depth)	+0.068
Bouguer reduction (completely rigid crust, or regional compensation extended to the whole earth)	+0.244

PRINCIPAL DEDUCTIONS AS TO ISOSTASY

As respects the broader geophysical problems, the facts revealed by gravity measures are that the crust of the earth as a whole is not completely rigid, but that it is in a state of equilibrium; that there is

not complete local isostasy, but that the compensation is regional to a moderate extent; and that smaller topographic features of moderate extent are not in equilibrium, but are, in part at least, supported by the strength of the crust and compensated as a part of the block on which they rest. Gravity measures indicate that the principal compensation effects lie probably within 100 miles of the station, both horizontally and vertically. They give no information at all, however, as to the vertical distribution of compensation, or the extreme depth to which it is sensible.

In the light of this and other evidence, a reasonable conception of isostasy is that the crust of the earth is a shell of appreciable thickness in general equilibrium, but of sufficient strength to partially maintain considerable surface irregularities; that this crust, probably without any definite lower limit, rests on a denser substratum which, owing to pressure and heat, is capable of horizontal displacement. When any area of this crust is sufficiently loaded it will bend like a flexible plate, or it will break, and in either case it will press down into the substratum. When it is unloaded it will rise, pushed up by the substratum. The transitions in compensation, both horizontally and vertically, will be gradual, excepting where ruptures in the crust occur owing to stresses beyond the breaking limit. This condition of isostasy acts continuously, and all deductions as to crustal formation and earth movements must be consistent with it. But isostasy is only a tendency toward equilibrium, which becomes active when equilibrium is disturbed. Isostasy will often appear to be obscured by other stresses and conditions in the earth's crust.

THE METAMORPHIC IRON FORMATION OF THE EASTERN MESABI RANGE, MINNESOTA, AND ITS RELATION TO THE EMBARRAS GRANITE

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ABSTRACT

The easternmost portion of the granite of the Mesabi Range, Minnesota, was regarded by C. K. Leith as intrusive in the Animikian iron formation, and mapped as Embarras granite. Obervations in the field, as well as microscopical and chemical examinations of the contact of both formations, convinced the writer that Leith's opinion accounts in a satisfactory way for the relation of granite and iron formation, but that the assumption of a pre-Animikian age of the granite is not supported by decisive facts and, besides, it leads to a very complicated explanation of the contact zone and its metamorphism. A detailed description of this contact zone is given in this paper. It is an unusual mineral association with some peculiar minerals, of which a blue-green amphibole has been discussed previously elsewhere.

INTRODUCTION

The granite north of the area of the Biwabik iron formation, the Giants Range granite, was regarded by the first geologists of the Minnesota Geological Survey as Archean, on which the Animikian iron formation had to rest unconformably. In the eastern part of the Mesabi Range, Charles K. Leith[1] separated from the older Giants Range granite a later granite as intrusive in the Animikian iron formation, and called it Embarras granite. In 1919, Grout and Broderick[2] rejected the interpretation of this Embarras granite as intrusive in the Animikian, and mapped it with the Lower-Middle Huronian Giants Range granite:

> In much of the east Mesabi . . . . the iron formation rests on the granite. These contacts do not show any intrusions of granite into the sediments. The granite had evidently been weathered before the sediments were deposited, for it now shows an upper zone (and a zone around many pebbles in the conglomerate) altered to a quartz-chlorite-garnet rock.

On Plate III, *A*, of the same bulletin a photograph is given with the legend: "Basal Conglomerate of the Biwabik Iron Formation,

[1] Charles K. Leith, "The Mesabi Iron-bearing District," *U.S. Geol. Surv., Mono. 43* (1903).

[2] Frank F. Grout and T. M. Broderick, "The Magnetite Deposits of the Eastern Mesabi Range, Minnesota," *Minn. Geol. Surv. Bull. 17* (1919), pp. 5, 7.

resting on the Granite. The Granite Mass and Boulders Show Dark Chloritic Borders."

This view, therefore, is opposed to that of Leith. The occurrence of a basal conglomerate with boulders of the underlying granite would be an especially conclusive argument against later intrusion. On the other hand, observations with regard to the stratigraphic relations of iron formation and granite and a peculiar mineral association at the contact of both, which urged Leith to assign a Keweenawan age to the granite,[1] cannot be waived so easily. This mineral association was recently described by Allison[2] as a "unique quartz-chlorite-amphibole-garnet-rock" and interpreted as a weathered residue of the granite metamorphosed first by regional and dynamical processes, and then by the contact action of the Keweenawan gabbro.

Some years ago the writer became interested in this problem and repeatedly has spent several days at a time at the very spot where Leith became convinced of the intrusive character of the granite and where Grout, Broderick, and Allison thought to find decisive facts in favor of a pre-Animikian age of the same granite. A careful microscopic examination of the granite and its contact zone, and also a number of chemical tests, followed the field work. In the course of these studies, four minerals not mentioned by former investigators were found in the contact zone, three of which were of frequent occurrence. Moreover, the detailed study of this border zone in its mineral composition and in its relation to the granite and the iron formation forced the writer to return to the hypothesis of intrusion. Consequently, this contact or border zone is to him a part of the iron formation, metamorphosed by, and partly mixed with, the intrusive granite, and not a weathered residue of the underlying granite, metamorphosed by some other processes.

THE GRANITE

The granite of the eastern Mesabi Range (Embarras granite of Leith) "outcropping principally in the core of the Giants Range for

[1] Charles K. Leith, *op. cit.*, pp. 160–64, 186–88.

[2] Ira S. Allison, "Weathered Granite Twice Metamorphosed," *Jour. Geol.*, Vol. XXXIV (1926), p. 281.

a distance of 12 miles southwest of Birch Lake"[1] is regarded by Allison as a late phase of the Lower-Middle Huronian Giants Range granite, but still of pre-Animikian age. To his description[2] a few points of interest may be added. The plagioclase is an acid oligoclase of Ab 85–90. Biotite occurs in large, well-defined, brown individuals, partly fresh, partly altered to chlorite with very low birefringence and abnormal interference colors (penninite). The by-products of this alteration are titanite and calcite and, in one instance, zoisite. Common green hornblende is present in about the same amount as biotite. In places, it is intergrown in parallel orientation with a colorless amphibole of higher birefringence. The difference amounts to several points in the third decimal place. In sections parallel to (010) both amphiboles alternate several times; the boundary lines are irregular but without transition.

CONTACT ZONE BETWEEN GRANITE AND IRON FORMATION

The contact zone between the granite and the iron formation can be studied to the best advantage southeast of the hill on which the little mining place named Babbitt is built, in the NW. $\frac{1}{4}$ of Sec. 17, T. 60 N., R. 12 W. Numerous exposures of this contact are visible in the woods; moreover, a gorge cuts through the iron formation into the granite. The contact line is invariably marked by a highly characteristic border. It is the border interpreted by Grout and Broderick[3] as a "weathered zone of the granite, altered to a quartz-chlorite-garnet rock." It was later described by Allison.

The results of the writer's study of the same border zone differ in some essential points from the statements of these geologists. The contact borders are, as a rule, very small; only in a few instances do they reach a little farther into the iron formation. Two varieties of very different mineral association can be observed, one characterized by the grouping of biotite-garnet, the other consisting of amphiboles and, in places, of pyroxene. A chloritic border has never been seen here by the writer.

[1] Ira S. Allison, "The Giants Range Batholith of Minnesota," *Jour. Geol.*, Vol. XXXIII (1925), p. 500.

[2] *Ibid.*, pp. 503–4.

[3] *Op. cit.*, p. 7.

BIOTITE-GARNET BORDER

In this first variety of mineral association well-bounded crystals of red garnet abound in a velvety aggregate of a dark-green mineral with perfect cleavage; large quartz grains of bluish tinge, as in the granite, are also abundant; and occasionally some pink feldspars are seen in the border zone. Under the microscope, the green mineral appears either in very fine scales without definite shape (sericite-like) or in well-defined blades or patches of uniform extinction. In a few instances, numerous blades of varying orientation are clustered; in other places, large patches of the mineral contain blades with sharp contour lines. The perfect cleavage reveals itself in those blades in the form of fine straight lines. The appearance is decidedly that of biotite. The optical properties point in the same direction. The pleochroism is strong, from deep-green to yellowish. The birefringence is high, at least 0.050, probably plus a few points in the third decimal place. The highest refractive index was found to be between 1.650 and 1.655. The mineral is practically uniaxial, negative. There are numerous zircon crystals included in the green mineral, with dark pleochroic haloes of a radius to 0.04 mm. Also, apatite has haloes which, however, are less dark and have a shorter radius—0.013 mm. In contact with magnetite the green color of the mineral changes to brown.

These optical properties are evidently those of biotite. The green color is somewhat unusual, although not altogether unknown in biotites, especially in metamorphic rocks; and the high percentage of ferrous iron in the contact border may account for it. Indices and birefringence agree well with biotite III given in an important publication of Grout,[1] which biotite contains 23.23 per cent FeO, 3.03 Fe_2O_3, and 9.24 per cent MgO. Allison classifies the green mineral in the border zone as stilpnomelane, without giving chemical tests or optical data, except high birefringence.[2] However, the birefringence of stilpnomelane is much higher in all varieties whose optical constants are given by Larsen,[3] and in the stilpnomelane de-

[1] Frank F. Grout, "Notes on Biotite," *Amer. Mineral.*, Vol. IX (1924), p. 161.

[2] *Jour. Geol.*, Vol. XXXIV, p. 282.

[3] Esper S. Larsen, "The Microscopic Determination of the Nonopaque Minerals," *U.S. Geol. Surv. Bull. 679* (1921).

scribed by Grout[1] the gamma index is only 1.615 ± 6 and the double refraction is 0.070, against 1.650 and 0.050, respectively, in the Babbitt green mineral.

Convinced that chemical analysis alone could decide, the writer sent a sample to Dr. Larsen. He writes about it: "I have had Mr. Gonyer, our chemist, make a rough determination of the alkalies in the purest part of the biotite rock. The determination gave K_2O 4.23 and Na_2O 0.99. This, it seemed to me, would confirm the determination of the mineral as biotite." A chemical test for water, made by the writer on a rock sample, consisting mainly of the green mineral, yielded 0.25 per cent water below 110° and 3.1 per cent above 110°. In stilpnomelane the water content is always higher, mostly much higher. Thus the chemical analysis confirms the observations under the microscope: The green mineral is a biotite although a peculiar variety of this mica.

Besides the mineral classified as stilpnomelane, Allison mentions also chlorite as an important constituent of the border zone. Unfortunately, no optical or chemical data are given. But Allison distinguishes this chlorite plainly from stilpnomelane and ascribes its origin to regional and dynamical metamorphism, while "the development of stilpnomelane may have been . . . . a contact effect."[2]

The writer was unable to find chlorite in the border zone, although quite a number of sections were studied which, in their general appearance, fit well in the description given by Allison. It is true that the green mineral frequently is developed in a sericitic manner and, in such cases, the high double refraction can only be seen with a high-power objective. With these precautions it became evident that the scaly form belongs to the same green biotite. Penninite occurs in the granite, and in the granitic constituents of the borders, as an alteration product; which is, however, subordinate, as far as observations were made, also in the granite in immediate contact with border zone.

The most conspicuous mineral in these borders is garnet, in perfect crystals of almandite in the usual combination—dodecahedron and trapezohedron (211)—from 1 to 3 mm. in diameter. It is optically

[1] Frank F. Grout, "Notes on Stilpnomelane," *Amer. Mineral.*, Vol. IX (1924), p. 209.

[2] *Op. cit.*, pp. 284–85.

isotropic. Inclusions of quartz are abundant. The green biotite is especially plentiful around garnets. Magnetite is scarce in the contact border, all the iron being used up for the formation of silicates.

Of special significance is the immediate contact between granite and the border zone. It is frequently observed that the minerals of the former stretch out into the latter. In such cases, the plagioclase in particular shows characteristic peculiarities. Where a plagioclase crystal enters the border zone, it becomes intergrown extensively with quartz, often in pegmatitic manner; several grains of quartz appear in the feldspar in uniform orientation. Furthermore, the plagioclase encloses numerous scales of the green biotite. One instance is especially instructive. A large plagioclase, in the granite free of inclusions and with well-developed lamellae, reaches out into the contact border. There it is filled with quartz and green biotite to such an extent that the feldspar soon becomes obliterated. But at some distance from the granite the plagioclase reappears, in the same orientation and with the same lamellae, which latter are a continuation of those observed in the granite. Besides quartz and biotite, also a few grains of plagioclase in different orientation are visible in the main individual; they have the same refraction but are without lamellae.

The writer can find but one explanation for these relations of granite and border: an iron-rich sediment, i.e., the Biwabik iron formation was intruded and metamorphosed by the granite. At the immediate contact, the granite, not yet wholly solidified, was mixed with the sediment; and the minerals resulting from this mixture crystallized simultaneously.

The plagioclase was frequently found to be more acid in the border than in the granite, containing 95 per cent albite, apparently on account of an addition of silica from the sediment. Such feldspars may be seen at a distance of a few inches from the granite almost completely filled with the minerals derived from the sediment and isolated from the granite by these minerals. Large quartz grains of a bluish tinge are found a little farther away from the granite. Also, these seem to be portions of the granite crystallized in the sediment. The final process was the formation of veins. These veins,

replacing orthoclase and plagioclase, form a mosaic of fine-grained quartz, green biotite, and rather large crystals of apatite.

The intrusion of the granite, suggested by the observations just described, finds a strong support in the occurrence of garnets in, and their restriction to, the contact zone. Indeed, this fact is so convincing that several experienced petrographers to whom I showed such specimens did not hesitate a moment in declaring these borders with garnet due to the contact action of the granite. There is no reason why regional or dynamical action, or why the intrusion of the gabbro at a considerable distance, should have accumulated such perfect crystals of garnet just in the immediate vicinity of the granite. The iron formation, in its varying composition, contains undoubtedly at many places sufficient aluminum for the formation of almandite.

AMPHIBOLE-PYROXENE BORDER

Quite different in composition, though not in its appearance, is the other variety of contact rims, containing amphiboles and some pyroxene instead of biotite. This border also forms a well-defined, dark green zone, accompanying the granite. It contains, besides the ferromagnesian minerals, large bluish quartz individuals, but only occasionally garnet. The most conspicuous mineral is the blue-green amphibole, described recently.[1] In some specimens grünerite is a constant companion of the blue-green amphibole. The grünerite forms independent, well-defined blades. Polysynthetic twinning is common, and the green tinge of c and a faint pleochroism are easily noticeable. In other specimens of the contact border grünerite is absent. The writer was unable to find the orthorhombic amphibole gedrite, mentioned by Allison.[2]

In one specimen the blue-green amphibole is associated with a diopsidic pyroxene. It appears in large pale-green plates without pleochroism, and contains numerous inclusions of the blue-green amphibole but no grünerite. The same pyroxene occurs about 1 mile due south of Babbitt in the old mining pits. The indices of

[1] Stephen Richarz, "A Peculiar Blue-Green Amphibole from the Metamorphic Iron Formation of the Eastern Mesabi Range, Minnesota, *Amer. Mineral.*, Vol. XV (1930), pp. 65–69.

[2] *Op. cit.*, p. 283.

refraction are as follows: $\alpha = 1.693$, $\beta = 1.700$, $\gamma = 1.720$, all ± 0.002. According to the diagram of Winchell,[1] these indices would correspond to a composition of approximately 40 per cent hedenbergite and 60 per cent diopside. The extinction angle is 40°–41°. The amphiboles associated with this pyroxene from the mining pits were described in a previous paper.[2] Both amphiboles there mentioned differ more or less from the blue-green amphibole near Babbitt.

Also, in this second variety of borders feldspar may be seen growing from the granite into the sediment, and the plagioclase is there intergrown with quartz and amphibole. The picture is exactly the same as in the biotite-garnet border. Apatite is common in these borders; green biotite seems to be absent; some brown biotite blades may belong to the granite. Quartz occurs in the border either in layers of a rather fine grain or in larger individuals. The stratified quartz, apparently, is a part of the sediment and is associated with amphiboles. The larger quartz grains, however, must be derived from the granite. They contain numerous inclusions of amphibole, which show that this quartz, as well as the plagioclase, was not brought into the sediment, but that the solution of the granite crystallized in this border, and that thus the newly formed quartz could encompass the minerals made from the sediment.

The border zones just described invariably accompany the granite, either in the biotite-garnet combination or in the amphibole-pyroxene grouping. It is, as far as the writer's observations go, always a narrow rim between granite and iron formation in which the granite minerals are mixed with those of the sediment. The effect of the granite, however, is noticeable beyond this rim; just how far, it is impossible to determine, because the distance of the scattered exposures from the granite is unknown. In one case, at a distance of 2 feet from the contact, blue-green amphibole is present with some apatite. At 4 feet, scaly biotite is seen in which fibrous, often radiating, grünerite is embedded; no blue-green amphibole is at that place. Quartz appears here partly in globular, fine-grained,

[1] A. N. Winchell, "Studies in the Pyroxene Group," *Amer. Jour. Sci.*, Vol. VI (1923), p. 510.

[2] Stephen Richarz, "Grünerite Rocks of the Lake Superior Region and their Origin," *Jour. Geol.*, Vol. XXXV (1927), pp. 699–700.

aggregates and partly in larger individuals with sharp edges. The former aggregates remind one of the original greenalite-chert rock; the latter are apparently of clastic origin.

In places, the zone of mixture of granite and sediment seems to be of greater thickness. H. V. Winchell[1] apparently observed such rocks west of Babbitt:

> Near the north end of Iron Lake the magnetitic quartzite undergoes queer metamorphosis. It still lies in nearly horizontal beds; but it gradually becomes less highly charged with magnetite, acquires feldspathic and quartzose material, and finally is changed to a regular crystalline rock, which is the syenite of the Giant's Range (i.e., Embarras granite). The lowest beds are the most perfectly crystalline.[2]

Still farther west Winchell reports from the bottom of a similar exposure an "indistinct gneissic structure which seems to lie nearly horizontal."[3] Presupposing a pre-Animikian age of the granite, Winchell interprets the intermediate zone as a fragmental formation of débris derived from the granite. In these places Winchell also speaks of a recomposed granite. "The difficulty is in establishing in the field the difference between a natural granite and a recomposed granite."[4] Winchell's observations admit of another interpretation. If the granite is intrusive, we have here a typical contact zone of the granite: on top a gradual transition of magnetitic quartzites into rocks in which the iron occurs in the form of iron-rich silicates; closer to the granite the zone of mixture as described from Babbitt, only thicker; and at the bottom, granite prevailing, with some minerals derived from the sediment, at one place even a *lit par lit* injection into the iron formation. A detailed re-examination of these places, and a study of the transitions with modern petrographic methods, should be of special interest and importance.

The serious difficulty inherent in all theories which regard the contact zone of granite and iron formation as a detrital or residual product of the granite is the high iron content of this zone which can in no way be derived from the granite. Allison found an iron

[1] H. V. Winchell, "Report of Field Observations Made During the Season of 1888 in the Iron Regions of Minnesota," *Minn. Geol. Surv., 17th Ann. Rept.*, (1889).

[2] *Ibid.*, p. 84. [3] *Ibid.*, p. 86.

[4] "The Geology of Minnesota," *Minn. Geol. Surv.*, Vol. IV (1900), p. 395.

oxide content of 19.52 per cent, mostly in the ferrous form.[1] Why the granitic residue should contain so much iron, not only here and there, but invariably, is an unsolved puzzle. If, however, this zone is interpreted as the metamorphosed iron formation, the iron content is a matter of course.

GRANITE CUTTING THROUGH THE IRON FORMATION

The conclusions based on the study of the contact borders can be confirmed by observations of the stratigraphic relations of the granite to the iron formation. It is true, these relations are not, at the first glance, as evident in the field as is usual with granitic intrusions. So it happened that the intrusion could be doubted and denied.

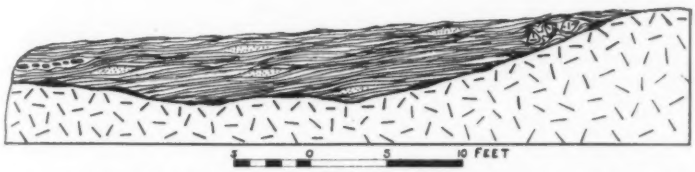

FIG. 1.—Wall of a gorge near Babbitt, showing the intrusive contact of the Embarras granite and the iron formation. The layers of the latter are cut by the granite.

However, a detailed examination of the stratigraphic conditions reestablishes, to my mind, beyond doubt, the intrusive character of the granite.

Figure 1 is representation of these conditions observed on the wall of the gorge in Sec. 17, southeast of Babbitt. This gorge, running SSW., lays bare the granite, and above it, the iron formation in a maximum thickness of $4\frac{1}{2}$ feet. The wall facing Babbitt is 40 feet long. The layers of the iron formation are slightly dipping toward SSW., almost parallel to the surface. The granite boundary is undulating and approaches the surface toward the NNE. end of the exposure, cutting there the strata at an angle of 30°. Before the granite reaches the surface, granite boulders seem to appear in the sediment, but a closer examination shows that the granite also lies behind the iron formation and that the "boulders" are protrusions of its irregular surface. A basal conglomerate with boulders of the underlying granite is not visible in this ideal exposure. The granite

[1] *Op. cit.*, p. 284.

is in contact with the various layers of the iron formation, which are cut off successively by the granite.

The same may be seen at many other places northwest of the gorge, where the contact frequently is exposed in the woods. Figure

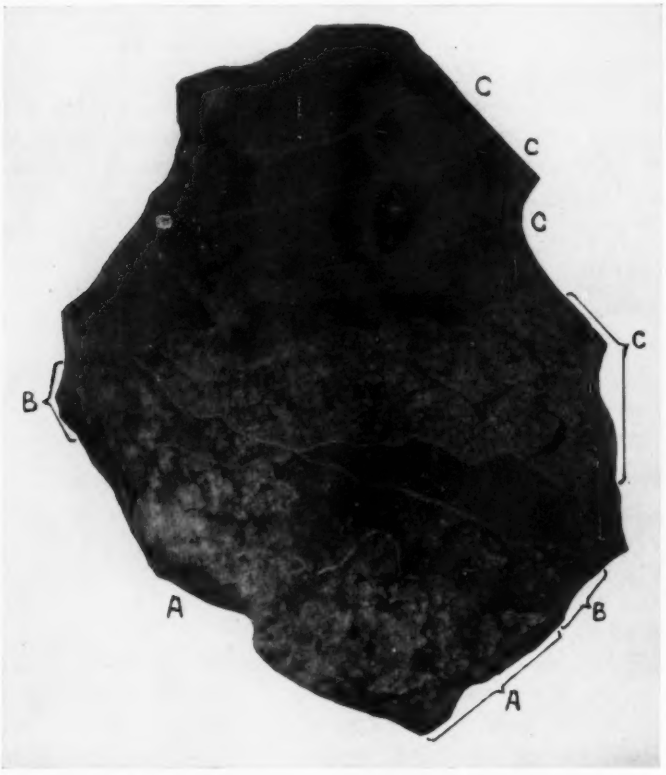

FIG. 2.—A specimen showing the contact of granite and the iron formation. The granite, *A*, cuts off the strata under an acute angle. The border zone, *B*, is in its outline parallel to the granite, but forms an acute angle with the strata of the iron formation. Quartzite layers, *C*, mark the stratification. The outlines, well visible in the specimen, are somewhat intensified in the photograph.

2 illustrates in detail these conditions. The stratification of the sediment is well marked by white quartzite layers. The granite

boundary forms an acute angle with the strata and cuts off a thick quartzite layer. The intrusive is bordered by a small contact rim, consisting, in this case, of quartz, amphiboles, and of minerals of the granite. The border zone follows the granite; and the outer outline of the border runs, like the granite outline, in an acute angle to the sediment. Such unconformities are the rule, whether the contact rim is characterized by the combination of biotite-garnet, or of amphibole-quartz. That is evidently not the appearance of a

FIG. 3.—A rock ledge southeast of Babbitt. The granite contains inclusions of the iron formation. Photo by Henry Retzek.

deposition contact. If sedimentation had taken place on such a flat granitic floor, one would expect to see the sediments winding around the irregular contours of the erosion surface of the granite and not a sharp cutting of the strata wherever they meet the granite.

In other instances portions of the iron formation are partially encompassed by granite and surrounded by a contact rim. Occasionally, small bands of the sediment may be seen deeply immerged in the granite and bordered on either side by the contact zone. Figure 3 is an illustration of these conditions.

THE "BASAL CONGLOMERATE"

After all these indications of a granite intrusion into the iron formation, the occurrence of a "basal conglomerate" containing boulders of the same granite would be highly surprising; and it cannot be accepted unless unmistakable evidence be given. Is such evidence established? The exposure, illustrated in Figure 1, fails to show such a basal conglomerate just where it ought to be. Grout and Broderick reproduce in their publication, on Plate III, *A*, a photograph which should be representative of what these authors interpreted as basal conglomerate. For the understanding of this picture it is essential to note that it is not the photograph of a more or less perpendicular wall, which would show granite boulders to a distance of some feet from the contact. It is the photograph of a slightly sloping rock ledge, as can be ascertained by closer inspection with a large reading glass. Thus the plane of the picture is almost parallel to the surface of the main granite mass. Such exposures are numerous southeast of Babbitt. At many places, the irregular and slowly rising surface of the granite pierces through the iron formation. Thus granite knobs of all possible shapes protrude out of the sediment. Quite a few of these knobs look exactly like boulders; and one would regard them as boulders, if it were not for the fact that their connection with the main granite mass beneath the iron formation can be established. Figures 3 and 4 are photographs of such outcrops. Figure 3 shows chiefly granite in which some portions of the iron formation seem to float. In Figure 4 the granite is exposed at the right and on top, and protrusions of it are seen in the adjacent iron formation. These protrusions become rarer toward the lower left, where the iron formation is thicker. The contact borders are well visible on the photographs. They follow invariably the contact between the granite knobs and the sediment. The steady occurrence of these borders around the knobs should be another proof of the intrusion of the granite. A weathered zone of such a regular shape around boulders would be, to say the least, highly improbable, while around the protrusions of an intruded granite such contact borders must be expected.

The problem of a basal conglomerate in the eastern Mesabi Range cannot be decided by reference to such surficial outcrops. Former

geologists of the Minnesota Geological Survey examined evidently more favorable exposures. Reference has already been made to Winchell, who studied extensively the area in question, especially the contact of granite and the iron formation. He does not mention a pronounced basal conglomerate, but at one place he observed that "the Animikie rock appears to be feldspathic in spots, as if there were boulders of syenite in it"[1] Such an indefinite statement cannot be regarded as decisive proof of the presence of a basal conglomerate.

FIG. 4.—A rock ledge southeast of Babbitt. At the right, chiefly granite with remnants of the iron formation; at the left, chiefly iron formation. Granite knobs appear in the iron formation. The dark rims around granite are contact borders. Photo by Henry Retzek.

Nor does the description of a "conglomerate" from the same place[2] bring more light on the problem. The occurrence of abundant hornblende with the granitic constituents points to conditions similar to those at Babbitt: an intrusion of granite in and a mixture of the intrusive with the iron formation.

On the other hand, it seems to be established that basal conglom-

[1] *Minn. Geol. Surv., 17th Ann. Rept.*, p. 83.

[2] "Geology of Minnesota," *Minn. Geol. Surv.*, Vol. V (1900), pp. 920–21.

erates occur farther west, where the iron formation and the Pokegama quartzite rest on the older Giants Range granite. Furthermore, conglomerate layers, or at least pebbles at various horizons of the iron formation, are not uncommon. They are referred to repeatedly in the publication of Grout and Broderick. Winchell mentions them from the top of the exposure north of Iron Lake.[1] They are seen also in the wall 1 foot above the granite represented by Figure 1 of this article. However, the pebbles of such conglomerates consist of quartz, as a rule, although one could not be surprised if some granite were found among them. The Giants Range granite forms, everywhere, the base of the iron formation; and, no doubt, this granite is the main source of the clastic constituents of this formation. Thus an aplite, seen by the writer at some distance from the underlying Embarras granite, might well be regarded, if it is a pebble, as a differentiate of the older granite.

To sum up, it may be said, that the opinion of Leith as to the post-Animikie age of the Embarras granite is well supported by many facts. At Babbitt one stands on the roof of the granitic batholith with its irregular surface protruding at places through the iron formation, and thus causing frequently the impression of granite boulders in the sediment. The granite surface rises toward the north until it forms the hill on which Babbit is built.

PROBABLE EXTENSION OF THE EMBARRAS GRANITE

According to the geologic map published with *Monograph 43* and reproduced in *Monograph 52*, the post-Animikian granite reaches eastward to Birch Lake, where it disappears. In this connection it seems not without interest that beyond Lake Superior, some 120 miles almost due south, a granite, intrusive in the Tylor slates and in the Keweenawan gabbro, was found by the writer to be of much greater importance than heretofore supposed. West of Mellen, Wisconsin,[2] the Tylor slate (younger than the Ironwood iron formation) is intruded extensively by granite, either in the form of dikes or in veins, at places forming injection gneisses with the

[1] *Op. cit.*, p. 86.

[2] *Bull. Geol. Soc. Amer.*, Vol. XXXIX (1928), p. 164.

slate. This zone is ½ mile wide across the strike. North of the slate Keweenawan gabbro is swarmed with granite dikes, in some of which big boulders of granite are floating. Toward the northeast the granite disappears, except for a few granite dikes. It reappears, however, farther north, first in the form of numerous dikes, the larger ones including gabbro. Finally the granite prevails. It is at the Power Dam of the Bad River, north of Mellen, where these relations can be studied in extensive exposures. In a recently published bulletin of the Wisconsin Geological Survey, the same granite is mapped by H. R. Aldrich as covering an area of more than 15 square miles.[1] It would be hazardous to connect the granites at Babbitt and those at Mellen and to ascribe them to one great batholith, although it seems to be possible. In any case, these two occurrences show that younger granites undoubtedly exist in the Lake Superior district and that the hypothesis "of two and only two periods of granitic irruption" (Laurentian and Algoman), recently repeated by Lawson[2] cannot be maintained. That the surface outcrops of these younger batholiths are not so extensive as those of the former periods is a matter of course.

On the map, the Embarras granite reaches almost 10 miles from Babbitt toward the west, and it is there in contact with the older Giants Range granite. Underground it seems to be traceable much farther west. According to Gruner,[3]

> during recent years drilling and underground mining have disclosed a number of intrusive dioritic masses near Aurora [intrusive in the iron formation]. . . . . There are indications that intrusions occur as far west as Virginia.[4] . . . .
>
> These veins and dikes in connection with acidic and basic intrusives in the Giants Range granite north of the iron formation suggest that possibly there is a large batholith younger than the Upper Huronian rocks beneath the Biwabik formation.[5]

[1] H. R Aldrich, "The Geology of the Gogebic Iron Range of Wisconsin," *Wis. Geol. Surv. Bull. 71* (1929).

[2] Andrew C. Lawson, "Some Huronian Problems," *Bull. Geol. Soc. Amer.*, Vol. XL (1929), p. 362.

[3] John W. Gruner, "Contribution to the Geology of the Mesabi Range, with Special Reference to the Magnetites of the Iron-bearing Formation West of Mesaba," *Minn. Geol. Surv. Bull. 19* (1924).

[4] *Ibid.*, p. 7. [5] *Ibid.*, p. 27.

SOME INSTANCES OF METAMORPHISM IN THE IRON FORMATION, POSSIBLY CAUSED BY THAT BATHOLITH

Such a batholith could account for the metamorphism of the iron formation east of Mesaba. As the writer pointed out in a previous paper,[1] grünerite was found in rocks of the Biwabik formation which are separated from the gabbro by non-metamorphic Virginia slates. It was suggested that an offshoot of the gabbro might have caused the metamorphism.[2] If there is a granitic batholith, more or less approaching the present surface at places, the presence of highly metamorphic rocks can no longer be surprising.

Only one find of grünerite northeast of Mesaba station could be mentioned in the foregoing publication. A search for more material was not without success. North-northwest of the town of Mesaba, in Sec. 15, T. 59 N., R. 14 W., there are some exposures in the woods. This place must be close to the shaft from which Grant[3] took the grünerite rock. Some of these rocks are rich in grünerite in well-developed blades with perceptible pleochroism, or the grünerite occurs in fine, radiating laths. It is associated with magnetite, quartz, calcite, some siderite, and with green amphibole, the latter strongly pleochroic, c being green with a faint bluish tinge, b peagreen, and a almost colorless, somewhat yellowish. However, c is not as blue as in the blue-green amphibole from Babbitt. Moreover, the color varies occasionally in the same individual; it may even disappear completely. The highest refractive index is about 1.66, considerably lower than that of the blue-green amphibole.

Besides these minerals another substance occurs in the rock which could not be identified as any known mineral. It forms aggregates of irregular shape and of indefinite, blurred extinction. The elongation, as far as can be ascertained, is parallel to c. The pleochroism is easily noticeable though not strong; c is sulphur yellow, a pale yellow to colorless. No cleavage is visible. The indices could not be determined exactly, but γ must be very close to 1.66; the birefringence is between 0.025 and 0.030; the negative axial angle $2E = 120°–130°$, i.e., $2V = 63°–67°$. A qualitative chemical test yielded, besides silica and much iron, some alumina, very little lime, only

[1] *Op. cit.*, p. 698. [2] *Ibid.*, p. 707.

[3] U. S. Grant, "Geology of Minnesota," *Minn. Geol. Surv.*, Vol. V (1900), pp. 364–65.

traces of magnesia, and no water at a red heat. The substance does not dissolve in boiling hydrochloric acid. The only interpretation I can give at present is that it might be a crystallized form of greenalite; however, in a higher metamorphosed state than the crystallized matter occurring farther west in greenalite globules, described at some length by the writer[1] and formerly regarded as amphibole. As a rule, needles and blades of grünerite pierce through the yellow substance, while the green amphibole occurs in clusters together with quartz.

The same rock was found along the railroad Mesaba-Babbitt, 1½ miles from Mesaba station (Sec. 22, T. 59 N., R. 14 W.) and in Sec. 11 of the same township, 150 feet before a branch of the railroad switches off to the Spring Mine. One-half mile south of this place, also at the railroad, a rock is exposed containing large pyroxene crystals. It is the same diopsidic pyroxene as was found near Babbitt. Pyroxene is associated with magnetite and quartz, but no grünerite is present.

Such instances of metamorphosed iron formation could undoubtedly be multiplied. Those mentioned are scattered over a few square miles, but it was impossible to connect the outcrops in the densely wooded country. Northeast of this area the iron formation was mined in the Spring Mine (open pit). No grünerite, nor any other mineral pointing to metamorphic action, could be found; and the chief iron ore is hematite. The same is true of the old pits at Mesaba, west and northwest of that area. To the south, the non-metamorphic Virginia slate occurs, which is, still farther south, metamorphosed by the gabbro. An approach of the hypothetical batholith to the surface could well account for all these irregularities in the metamorphism of the iron formation, while the assumption of a deep burial would not explain the immediate vicinity of metamorphosed and non-metamorphosed portions.

The actual irregularity of the batholith is well illustrated at Babbitt, where its undulating surface has led to the assumption of an erosion of considerable relief. At places, the iron formation is only a thin veneer through which the granite is seen. If this veneer were a little thicker, the granite would be hidden altogether and one

[1] *Op. cit.*, pp. 701–4.

would be highly puzzled by the high metamorphism of the sediments. How much of this metamorphism is due to the gabbro, which reaches the surface south of the iron formation, is not known up to date. In the immediate contact with the gabbro, fayalite and hypersthene associated with grünerite, seem to be the characteristic contact minerals. At the easternmost end of the range, granite and gabbro come close together, and there the combined metamorphism of both must have been active.

The writer is very much indebted to Dr. Larsen and his analyst, Mr. F. A. Gonyer, for the partial analysis of the biotite, and to Dr. Johannsen for the reading of the manuscript; also, to his friend Henry Retzek, who assisted him in the field work and made a special trip to Babbitt to photograph characteristic exposures.

STREAM HISTORY IN AND ABOUT KICKING HORSE PASS

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ABSTRACT

It is observed that there are abandoned waterfalls and other evidence of the former presence of a stream in the now dry, eastern end of Kicking Horse Pass. These conditions are attributed to a westward tilting or elevation of the region to the east in response to the removal of the ice load after the close of the Pleistocene glaciation or to tectonic uplift of recent date in the eastern Rocky Mountains of Canada.

INTRODUCTION

Location.—This paper is a preliminary discussion of certain observed features of stream history in the region of Kicking Horse Pass and the development of the pass itself, which is a wind gap not formed by stream piracy. The area covers about eighty square miles, located partly in Alberta and partly in British Columbia along the line of the Canadian Pacific Railway. The Continental Divide crosses the region in a northwesterly direction.

Drainage.—To the east of the Continental Divide, the master stream, Bow River, flows southeastward; and Bath Creek, an important tributary, joins it from the northwest and drains Bath and Waputik glaciers. On the west side of the divide, Kicking Horse River runs westward to the Columbia. The Kicking Horse is joined about five miles west of the divide by Yoho River from the north. Prior to this union, the Kicking Horse flows through a steep-sided canyon. After combining with the Yoho, it follows a much more open valley with extensive gravel flats.

PHYSIOGRAPHIC HISTORY

Kicking Horse Pass.—To an observer crossing the Continental Divide by way of the railway or the highway through Kicking Horse Pass, an extraordinary feature is apparent. As one enters the pass from the east he sees, facing eastward, abandoned waterfalls, dry rapids, and other features indicative of the former presence of a stream which, not long ago geologically, flowed eastward, but has

since disappeared. There is no evidence of piracy or other event in normal river history that would divert such a stream and change the pass to a wind gap. Therefore, some other cause for this condition must be sought. It is suggested that a westward tilting or rising to the east of the land might be responsible for this drainage change. Let us consider the history of some of the streams in order to discover what light they may throw upon the matter.

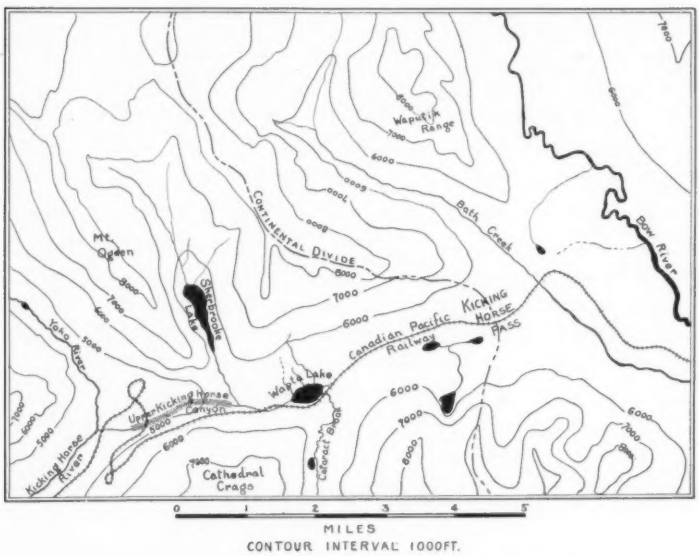

FIG. 1.—Sketch map of region about Kicking Horse Pass

Bath Creek.—Today Bath Creek is a misfit stream as it occupies, especially in its lower reaches, a rather broad valley. Below the point where the creek passes the eastern entrance to Kicking Horse Pass, its channel becomes suddenly broader and its valley more mature than above that place. Immediately south of the southern tip of the Waputik Range there is a tiny lake emptying in an easterly direction into Bow River. A small, intermittent stream just south of this lake and its outlet also enters Bow River. It requires no great stretch of the imagination to suppose that the region to the

east was once slightly lower, possibly from tilting in that direction. This could divert the upper waters of Bath Creek eastward into Bow River around the tip of the Waputik Range by way of the valley now occupied by the small lake and its outlet, or the intermittent stream south of these. This should behead Bath Creek and leave the lower waters as a small tributary to Bow River. The same tilting should move the Continental Divide westward and so cause drainage to pass from the west through the Kicking Horse Pass. Such a stream, running out from the pass, might join the beheaded, lower part of Bath Creek. This condition may have existed soon after the departure of the Pleistocene ice from the Rockies, and must have obtained for some time, with the production of the broad channel of the lower part of Bath Creek and the now abandoned rapids and falls in the eastern side of Kicking Horse Pass. Rising of the land to the east (or tilting westward) eventually would change the drainage to its present lines by uniting the upper and lower parts of Bath Creek and diverting drainage through the pass westward down the Kicking Horse.

Yoho and Kicking Horse rivers.—Wapta Lake lies just west of the Continental Divide and is the source from which Kicking Horse River rises and flows westward into its canyon. If the original, eastward tilting suggested in the last paragraph took place, it would require only a relatively small movement to divert the outflow from Wapta Lake eastward through Kicking Horse Pass. With it would go also the waters of Cataract Brook and Sherbrooke Lake, both of which now empty to the west. The former, Cataract Brook, flows north and then west into Wapta Lake; the outlet of Sherbrooke Lake forms a barbed tributary to the Kicking Horse just at the head of its very youthful canyon.

The valleys of Yoho and of *lower* Kicking Horse rivers are fairly wide and more mature than the canyon of *upper* Kicking Horse River. This part of the Kicking Horse is a very young tributary to the main Yoho–Kicking Horse system, and appears to have been acquired at some rather recent date. The original tilting to the east we saw could divert Wapta Lake and its feeders through the pass and cut off their waters from the Yoho–Kicking Horse drainage basin. If, later, the land to the east rose and the Continental Divide

migrated eastward, Wapta Lake may temporarily have drained both east and west after the manner of Two Ocean Lake today, in the Yellowstone National Park. Gradually the westward drainage won at the expense of the eastward. The swampy region at the top of the pass and the abandoned channel features in its eastern entrance are indications of the very recent date at which the diversion occurred. The supposition that stream piracy by upper Kicking Horse River took place to divert westward the waters of Sherbrooke and Wapta lakes through tapping, is a remote possibility; since it is not reasonable to suppose that that stream cut its canyon *headward*.

Time of change of drainage.—The time of this change of the drainage cannot have been long ago, geologically. It is certainly postglacial, since many of the streams cut through glacial deposits and occupy glaciated valleys. The extreme freshness of the abandoned waterfalls and rapids in Kicking Horse Pass points to a very late time of abandonment and suggests, when their degree of preservation is noted in conjunction with the size and depth of Kicking Horse Canyon, that a stream may have continued to run eastward through the pass even after the westward drainage through the upper Kicking Horse was well established. The inference is, of course, that tilting was gradual.

CAUSES OF CHANGES IN DRAINAGE

Two possible causes for the tilting and consequent change of drainage in the Kicking Horse Pass region are suggested. The first of these is adjustment of the crust of the earth after the removal of the Pleistocene ice; the second explanation is linked with recent tectonic movement in the eastern Rocky Mountains. Let us consider these two possibilities.

Postglacial crustal adjustment.—During the Pleistocene epoch the Rocky Mountains of Canada were covered by a large and thick ice sheet.[1] Subsidence of the crust under the weight of this ice is assumed to have gone on. Rise of the crust after the final retreat of the ice is a phenomenon reported in certain places. If such an ice cap or sheet were heavy enough, depression of the underlying and immediately surrounding region might be expected. With the retreat

[1] R. A. Daly, "Geology of the North American Cordillera and the Forty-ninth Parallel," *Can. Dept. of Mines, Geol. Surv., Memoir 38* (1912).

of the ice at the close of the Pleistocene, adjustment of the land surface to the new conditions followed gradually. By analogy with the raised beaches of the New England coast, this adjustment was not instantaneous. In New England adjustment was so slow that beaches and wave-cut terraces had time to form and ultimately be carried well above the present sea-level. That movement in the Canadian Rockies was gradual is deduced from the seemingly slow abandonment of the Kicking Horse Pass by its stream and the simultaneous and later cutting of Kicking Horse Canyon. Such a rise or tilting may have been due to isostatic adjustment of subcrustal material or merely to the elastic resilience of the crust itself regaining its former elevation after the ice had vanished. It is beyond the scope of the present paper to attempt to determine which of these possibilities is more likely.

Recent tectonic uplift.—It is possible that the tilting or uplift in recent times in the Canadian Rockies, which produced the changes noted in the drainage, is of tectonic origin. Evidence of recent movement of this kind has been described[1] thus:

> A recent movement in the region of the Great Plains seems also to be suggested by certain physiographic features. Extensive tracts in central Kansas and Nebraska bear an aspect of pronounced topographic youth, suggesting that they have been lying, until recently, near the neutral horizon between erosion and deposition, and have lately been raised on the western side. In the Dakotas, there are broad gradation plains of abandoned river-courses which cross the present valley of the Missouri River. Their present gradients, and their elevation above the present river-bottoms of the region, also imply a westward elevation. These and collateral phenomena, taken with the remarkable movement of the Keewatin ice-sheet from what is now the lower to what is now the higher side of the plains, seem best satisfied by the view that until the close of the Glacial period the western side of the Great Plains was lower than now, or the eastern side higher than now, relative to the common surface-level. . . . . On the western side of the continent there is much evidence of recent movement, some of which appears to have taken place since the close of the Glacial period, as usually defined.

In other words, there may have been a tectonic uplift of the western part of the Great Plains and of the eastern Rockies. Such a movement should affect the drainage in the postulated manner as readily as would an uplift due to adjustment after deglaciation.

[1] T. C. Chamberlin and R. D. Salisbury, "Geology," Vol. III (1906), pp. 518–19.

CONCLUSIONS AND SUMMARY

Whatever caused stream diversion and adjustment, and there may have been a combination of factors, the results are alike. At some time before the establishment of the present elevation, topographic features developed unlike those now seen. Then, lower Bath Creek may have headed in Kicking Horse Pass and flowed southeastward into Bow River. The Continental Divide lay somewhat west of its present position, possibly at or near Wapta Lake, or in the region of the present Kicking Horse Canyon. Cataract Brook flowed in from the south, and an east-flowing stream connected Wapta and Sherbrooke lakes. The present valley of Bath Creek was probably unoccupied from the southern end of the Waputik Range to a point opposite Kicking Horse Pass. Its present headwaters then drained eastward around the end of the Waputik Range and entered Bow River through still recognizable channels. Yoho River flowed much as now, south and then southwestward to join the main or lower Kicking Horse. This channel lacked any large tributary where now the upper Kicking Horse leaves its canyon.

Adjustment came with westward tilting or eastern uplift. It came after the ice was gone and the valleys had been exposed to stream action long enough to establish the drainage just outlined. The Continental Divide migrated eastward. The flatness of Kicking Horse Pass for two or three miles is such that it would require but a small tilting to upset its drainage. The size of Kicking Horse Canyon implies a fair interval of cutting. Evidently the tilting was rather slow with gradual abandonment of the pass. The cause of the movement is not certain, nor can it be finally ascertained at this time, although future work may give more clues. Crustal adjustment to loss of load in postglacial times, or recent tectonic uplift of the eastern Canadian Rockies, are suggested as possible explanations. It seems strange that the Rockies should rise more than did the more heavily glaciated Purcell and Selkirk ranges to the west. The reason for a tectonic uplift is likewise not easily understood. Adjustment to relief of ice-load in the western plains seems doubtful, since there the sheet was relatively thin. The question is not yet settled as to the cause; the results, however, are clear.

PYRAMIDAL JOINTING IN SHALES

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ABSTRACT

A horizontal band of pyramidal joint planes has been found in thick shales. This band underlies a layer of large concretions which were evidently a factor in the formation of the joints. The joints make angles of about 45° with the vertical. These angles appear not to have been modified by friction. This is explained by a lack of pressure normal to the breaking planes. The significance of these joints lies in their bearing on the topic of normal pressures. Normal pressures seem to have been incompletely considered in the geological literature of fracturing, yet they appear to be an important consideration in the formation of joints.

DESCRIPTION

These joints (Fig. 1) are in Hamilton (Moscow) shales south of Crowbar Point on Cayuga Lake. Joints set at similar angles are common in these shales, but as a rule they are not spaced regularly enough to give this conspicuous pyramidal effect. The band is about 3 feet high, is parallel to the bedding, and is exposed for a distance of 20 feet.

A few joints in the band are nearly vertical. The prismatic and pyramidal forms are distinct. Fourteen readings on the pyramid faces gave dips of 33° to 60°. Their average is 49°. The variation of this average from 45° is too small, probably, to justify deductions. If any conclusion is to be drawn therefrom, it is that there is a slight tendency toward an acute angle cutting the vertical axis of stress and an obtuse angle cutting the horizontal axis. Although the individual joints are both steeper and flatter than 45°, out of the fourteen readings only four have dips of less than 45°. Most of the oblique joints are arranged so that the rock breaks up into pyramids and often into bipyramids. The bounding planes of the pyramids mostly intersect sharply, but some curve around the corners in a continuous surface as shown in the lower center of the band in Figure 1. Some of the pyramids are remarkably symmetrical. They have from four to eight sides, but the higher numbers are not common.

ANGLE OF FRACTURE

Considering that friction noticeably modifies the angle that fault planes make with the 45° plane, and that the modification is especially great in faults in the local shales, it seems necessary to seek a reason for the fact that the pyramidal fractures remain so close to the angle of 45°. Friction appears not to have been effective.

FIG. 1.—Photograph of a band of pyramidal joint planes. They were formed between two layers of concretions in shales. The hammer lies on a concretion.

Perhaps the most obvious suggestion is that there may have been little or no pressure at right angles to the greatest axis of stress. This is analogous to the conditions existing when rocks are fractured in a testing machine. If cubes are crushed, pyramids are likely to be formed, and the fractures are near the angle of 45°. Investigation shows, however, that the field conditions are favorable to a more general theory, dealing not only with zero lateral stress, but also with lateral tensions lying in the range between zero and a tension

equal to the compressional stress. The literature of mechanics shows that, if a body is acted upon by two equal stresses at right angles to each other, one of which is positive (tensional) and the other negative (compressional), the planes of maximum tangential stress make angles of 45° with the stresses and the component of stress normal to the 45° planes is zero. If the stresses in the shales can be shown to satisfy these conditions, approximately, an explanation of the joints is at hand.

STRESSES IN THE SHALES

The only obvious reason for the marked development of joints in this band is the presence of two layers of concretions. The upper limit of the band is at a layer of large concretions, many of which are between 1 and 2 feet in diameter. One of these is marked by the hammer in Figure 1. The lower limit is at or near a layer of flat, much smaller concretions. That the lower limit of the band is not so even as the upper limit probably is due to the fact that the lower concretions were less effective in modifying the stresses that caused the fracturing. The upper concretions are of the same order of size as the pyramids.

The joints occur where the beds dip about 3°–4° S. on the limb of a low fold. It is difficult to see how the concretions would have played such an effective part if the joints were due directly to the horizontal regional stress active here during the Appalachian folding. The concretions do not lie across the path of horizontal forces, but are distributed parallel to the bedding. They are separated from each other and consequently are not as capable of carrying horizontal thrusts as a continuous competent bed. It seems more likely that the conspicuous development of these joints was dependent upon the modification of vertical forces by the concretions. The vertical forces were also at right angles to the greatest extent of the jointed section. It is exposed as a band, but doubtless it extends horizontally into the cliff as well as along the face of the cliff.

There are two possible sources of unusual vertical pressure across this jointed band. One is pressure from the growth of the concretions. If this was responsible, however, the pyramidal joints must antedate the major joints in the same rocks. The major joints in this district are known to have been formed after the growth of the

numerous concretions as both the dip and strike joints cut cleanly through concretions and a third set, the northwest quadrant joints, pass around the concretions. On the other hand, if the pyramidal joints were formed early, while the clay was still unconsolidated, the coefficient of friction may have been lower, and the lower the coefficient of friction the more nearly fractures approach the angle of 45°, according to the tables of Blamey Stevens.

The second possible source of pressure is that which causes the thinning of incompetent beds on the flanks of a fold. The jointed section lies not far below the competent Tully limestone. As the limestone folded it may have altered the vertical pressure on the underlying shale. The concretions lay across the path of such a pressure. Less compressible than the shales, it seems reasonable that they would cause a greater pressure under the center of a concretion than at the sides and between concretions, and hence cause an unusual amount of fracturing. Beyond the limits of the photograph, where the joints are not so numerous, oblique joints occur closely associated with two large concretions.

FIG. 2

In either case, whether the unusual conditions were due to the growth of the concretions, or to the influence of the concretions on forces passing through them, the result would be a tendency to squeeze the clay or shale sidewise from beneath concretions. The small concretions in the lower layer probably aided in the squeezing process. Figure 2 represents diagrammatically the forces acting on particles of the incompetent material, one pair of forces tending to move the particles toward a center and the other to move them away from the center. This seems to be all that is necessary in order to apply mathematical formulas pertaining to combined positive and negative (tensional and compressive) stresses. Forces are of one sign or the other accordingly as they tend to separate the bodies in contact at any surface, or tend to push the two bodies together.[1] The main forces need not be at right angles as they can be resolved into components at right angles to each other.

It does not seem necessary that one of the forces should be an

[1] Henry Crew, *The Principles of Mechanics* (New York, 1908), p. 244.

active tension of the type found on the convex side of an anticline. These pyramidal joints are located at an unusually steep dip on the steepest local fold, but the fold is low and it is unlikely that tension of that type was strong. Pressing concretions into incompetent material appears to be analogous in principle to pressing a rigid object down into mud. The force exerted may be mainly vertical, but the mud will be moved sidewise as a result of its incompetence.

If the incompetent material was pressed outward from beneath concretions, the lateral stresses analogous to tension would have had values somewhere in the range between zero and equality with the active vertical stress. This is a range in which the mechanical conditions are different from those in which both stresses are compressional, or in which one is compressional and the other tensional, but the tensional is the larger.

THE MECHANICAL THEORY

There are various mathematical methods of studying stresses to determine where fracturing will occur. The method that seems most usable in connection with these joints is one often found in the literature. It begins with a consideration of the plane of maximum tangential stress when the body is affected by a force acting in only one direction. This gives the well-known formulas:

$$P_t = P \sin \theta \cos \theta$$
$$P_n = P \cos^2 \theta$$

where P_t is the tangential component of pressure and P_n is the normal component. Since $\sin \theta \cos \theta$ is a maximum at $45°$, this is the theoretical position for breaking planes. Normal pressure is present, however, and is used in friction. This friction lowers the angle which the break makes with the direction of the force. If the pyramidal fractures were to be interpreted like ordinary faults their angle should be modified by friction.

A body may be acted upon by forces in two or three directions at right angles to each other. These forces may be all compressional, or all tensional, or a combination of the two. The formulas usually given are based on positive forces. If the body is acted upon by two

such forces,[1] the formulas for the total normal and tangential stresses on a plane at right angles to the forces are:

$$\begin{aligned}\text{total normal stress} &= P\cos^2\theta + Q\sin^2\theta\\ \text{total tangential stress} &= P\sin\theta\cos\theta - Q\sin\theta\cos\theta\\ &= (P-Q)\sin\theta\cos\theta\end{aligned}$$

If, however, the stresses P and Q are of unlike sign, as in Figure 3, the formulas are:

$$\begin{aligned}\text{total normal stress} &= P\cos^2\theta - Q\sin^2\theta\\ \text{total tangential stress} &= P\sin\theta\cos\theta + Q\sin\theta\cos\theta\\ &= (P+Q)\sin\theta\cos\theta\end{aligned}$$

In Figure 3 the compressional stress is called positive, reversing the usage in mechanics, because the dominant stress in the formation of the joints was compressive. This does not invalidate the formulas.

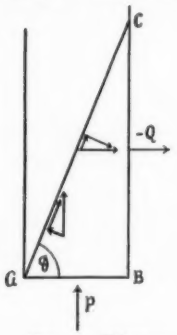

FIG. 3.—Diagram showing how the tangential stresses add to each other and the normal stresses tend to cancel each other when a body is acted upon by two stresses one of which is compressional and the other tensional, operating at right angles.

The greatest tangential stress is again at 45°. If P and Q are equal, but unlike in sign, the tangential pressure is twice as great as if there were only one active force. The directions of the arrows along the plane AC in Figure 3 show how the tangential stresses add to each other. The normal stresses point in opposite directions and tend to cancel each other. If P and Q are equal numerically, the normal pressure on the 45° plane is zero. If P and Q are not equal, the plane of no normal pressure varies from 45°. If the cohesion of the material is low, it seems likely that joints tend to form near the angle of no normal stress even when this angle varies somewhat from 45°.

Thus, under stresses of unlike sign, there is a tendency to form fractures against which there is little or no normal pressure. They would be formed earlier than faults that are due to a combination of compressive stresses. Under equal compressional and tensional stresses at right angles, the planes of greatest tangential stress

[1] G. F. Becker, "Finite Homogeneous Strain, Flow and Rupture of Rocks," *Bull. Geol. Soc. Amer.*, Vol. IV (1893), pp. 36–37; H. Crew, *op. cit.*, pp. 253–54; B. Stevens, "The Laws of Fissures," *Trans. Amer. Inst. Mining Eng.*, Vol. XL (1909), pp. 475–77.

and zero normal stress coincide at 45°. Perhaps such perfectly balanced conditions rarely exist geologically and a large tension is likely to result in fractures at right angles to the tension, not at an oblique angle, but when oblique fractures are formed these principles are active. If the pressures are both compressional, or both tensional, the tangential stresses are diminished and the normal stresses increased, but if one is compressional and the other tensional the tangential stress is increased and the normal stress diminished, thus facilitating fracturing.

Normal stresses appear to have been incompletely considered in geological literature. Although if the ratio of the compressional stress to the tensional stress is high, fractures may form at nearly the same angle as if both stresses were compressional, there are indications that, if the ratio is small, probably between $P = -Q$ and $P = -3Q$, fractures may be formed at a range of angles similar to those observed in the pyramidal joints, provided the cohesion of the rock is small.

THE PYRAMIDS

It is probable that the vertical pressure on the rock under a concretion was applied symmetrically, that it was greatest over a small area and diminished symmetrically toward the periphery, causing a symmetrical tendency for the incompetent rock to move away, and causing fractures symmetrical about a vertical axis. Thus the stress condition was that of an odd axis with the stresses nearly equal in the plane of the other axes. The well-formed pyramids point toward a vertical odd axis. They might be expected to require that the stress conditions be much the same in various directions at right angles to the axis of the pyramids. It seems more likely that various horizontal stresses would be alike and the vertical the odd stress in sedimentary rocks than that the vertical stress and certain horizontal stresses should be alike and the odd stress horizontal.

Under perfect conditions the fracture surface should be a cone[1] whose axis is parallel to the odd axis of stress. Theoretically, as soon

[1] B. Stevens, *op. cit.*, pp. 487–88; W. H. Bucher, "The Mechanical Interpretation of Joints," *Jour. Geol.*, Vol. XXVIII (1920), p. 714; T. A. Link, "Some Applications of the Strain Ellipsoid," *Bull. Amer. Assoc. Petroleum Geologists*, Vol. XIII (1929), p. 1456, Fig. 3; Vol. XIV (1930), pp. 240–41, Fig. 1.

as one of the axes becomes least, the fractures take the position of ordinary faults, forming two planes intersecting parallel to the intermediate axis. Practically, however, there may be an intermediate stage, like the pyramids formed in many crushing machine tests and in some of Daubrée's experiments. It seems probable that the pyramidal form rather than the conical form occurred in the rocks because cones cannot be packed together without interspaces but pyramids can. As the usual number of sides was four, and seldom more than six, these forms would pack together. The larger the number of sides, however, the nearer the approach to a cone.

CONCLUSION

These pyramidal joints show that such fractures can and do occur at or near an angle of 45° with the axes of stress. Probably they require complex special conditions, but the formulas used in explaining them call attention to the general principle that fracturing is easier if one of the stresses is tensional and one compressional than if both the greatest and least stresses are compressional. The pyramidal joints were fractures on which there was little or no friction. There seems to be a tendency to speak of fractures that are open, or tend to open as joints, as contrasted with slickensided faults. This seems to have a basis in fact. A study has been made of the joint planes of the Cayuga Lake district. Of six types of joints in the stratified rocks, all are interpreted as fractures against which there was little or no normal pressure at the time they were formed. There was tangential stress along most of them, but circumstances were such that there was little or no slipping.

ORIGIN AND GROWTH OF THE GREAT SALT LAKE OÖLITES

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ABSTRACT

After making a careful study of the Great Salt Lake oölites, the author concludes that the oölites of the zonal type originate at the water's edge, are washed upon the mud flats, and grow as they are driven inland by the wind. The laminae correspond to the seasonal changes of the area, and are the result of direct precipitation of amorphous aragonite from the evaporation of capillary water. The growth occurs during the early summer months when the temperature rise is greatest.

INTRODUCTION

Since the days of Rothpletz, who visited the region before 1892, most authors have concluded that the Great Salt Lake oölites were formed by organisms, i.e., certain algae belonging to the genera *Gleocapsa* and *Gleotheca*. Recent investigations, however, show the origin to be physico-chemical in nature rather than biochemical.

PHYSIOGRAPHIC CONDITIONS OF THE GREAT SALT LAKE AREA

The geographic location of Great Salt Lake is ideal for the origin and growth of the zonal type of oölites. This type may or may not possess radial structure, but is always composed of concentric laminations.

The lake, which has an elevation of over 4,100 feet, is situated in the temperate zone and is surrounded by mountain ranges. The climate is arid, consisting of a single major rainy period lasting from late September to May, and a short, intensely hot, dry summer. Many times during the summer season the relative humidity drops below 25 per cent. The average annual rainfall is about 20 inches. During the summer months the evaporation is 45 inches from a free surface.[1] As a result, the lake surface is frequently lowered more than 3 feet, leaving a low, flat, soggy, muddy shore exposed for some distance. During the early part of the summer season, strong variable winds

[1] J. B. Kincer, "Precipitation and Humidity," *Atlas of Amer. Agriculture*, Part II. "Climate," Advance Sheet No. 5 (1922), p. 48.

sweep across the lake and keep the water in constant agitation. At times the aëration is intense—often the spray is carried landward for a distance of thirty miles.

ORIGIN OF THE GREAT SALT LAKE OÖLITES

It is at the margin of the lake where the water is aërated most that oölites originate. Under such conditions, the lake water gives up

TABLE I

COMPOSITION OF THE NUCLEI OF OÖLITES TAKEN FROM DIFFERENT PLACES ALONG THE SHORE OF GREAT SALT LAKE, UTAH: THIS TABLE SHOWS A PREDOMINANCE OF THE LIGHTER MINERALS

Slide	Glass	Quartz Grain	Mica Flake	Cinder	Copper	Calcite Grain	Soot	Biotite	Opaque Mass	Limestone	Cerasite	Coal	Feldspar	Olivine
A	4	4	10	1	....	....	1	1	4	3	1	2	8	1
B	5	10	6	1	....	....	2	....	11	1	....	8	11	1
C	2	1	3	....	....	....	....	....	....	....	....	....	2	....
D	3	8	12	....	....	1	5	1	5	....	....	2	13	....
E	....	3	3	....	....	....	1	....	6	5	....	4	3	....
F	2	2	3	....	....	1	1	....	1	2	....	1	2	....
G	7	10	9	2	1	2	16	....	9	4	....	9	10	....
H	1	2	7	....	....	....	8	....	5	....	....	7	11	....
I	1	6	7	1	....	1	4	....	11	1	....	10	6	....
J	2	6	11	....	....	3	10	1	6	2	....	6	18	1
K	2	3	5	....	....	2	10	1	6	6	....	6	10	....
L	1	7	9	1	....	4	11	1	11	3	....	6	23	....
M	1	4	3	....	....	....	1	....	3	....	....	....	5	....
	31	66	88	6	1	14	70	5	78	27	1	61	122	3

some of its CO_2 and the resulting calcium carbonate precipitate collects about some foreign particles as nuclei to produce the oölites. Most Great Salt Lake oölites are formed around some solid nucleus, as is shown in Table I. This table shows that, of the 574 oölites examined, only four could have been formed about an alga or a gas bubble.

In examining the development of the Florida oölites, Vaughan[1] discovered that the mud of Marquesas Lagoon gives off large quantities of gas in which H_2S is important. After collecting samples and

[1] W. T. Vaughan, "Studies of the Geology and of the Madreporaria of Bahamas and Southern Florida," *Carnegie Institute of Washington Year Book No. 11* (1912), p. 158.

subjecting them to laboratory control, he noticed that "gas bubbles had formed in samples that had been kept wet" and that "small particles of calcium carbonate collect around the peripheries of the bubbles and form crusts. The crusts simulate oölite crusts." In some instances the "included gas diffused outward through the surrounding crust without rupturing it, permitting a contraction of the coating without losing the spheroidal form." This phenomenon is physicol-chemical in nature and is biochemical only in so far as the complex gas containing H_2S is the result of organic action.

The Great Salt Lake oölites are developed on mud flats adjacent to the shore, where large quantities of gas containing H_2S are present. A photomicrograph (see Pl. I, Fig. 1) of a sample collected from the water's edge shows nuclei with a single zone of amorphous calcium carbonate to be the dominant characteristic. Everywhere about the lake and the islands in the lake the shores adjacent to the mud flats contain oölites of this nature.

Oölites found at the bottom of the lake are very small, consisting of one or two laminae (see Pl. I, Fig. 6.) However, in areas where there are strong reverse winds, the oölites vary greatly in size with the small forms predominating. The larger forms have been blown into the water from the adjacent shore.

GROWTH OF OÖLITES

The oölites that settle to the bottom of the lake mix with the débris and precipitates, and, if preserved, form a very fine-grained deposit. Little, if any, growth takes place in the water.

The young oölites that are washed upon the flat, soggy shores are coated with a thin film of capillary water containing soluble salts; they are subjected to the various gases including H_2S; they are rolled about by the wind, rounded, pitted, and polished; they are dampened by lake water; they are subjected to strong heat which causes intense evaporation; and they are coated with the dust particles which settle from the smoky atmosphere. The growing process continues until the oölites are finally carried or rolled inland and deposited into dunes, or scattered over the higher terrains beyond the influence of the calcium bicarbonate-bearing water. By this time they have reached their maturity.

EXPLANATION OF PLATE I

FIG. 1.—Young oölites (×90) showing a single zone of amorphous calcium carbonate scattered among the débris of the shore. Taken from the water's edge on the west shore of Stansbury Island.

FIGS. 2, 8, AND 9.—Sample of oölites taken from the bottom of the lake near Saltair, where strong reverse winds occur. (×90)

FIG. 3.—Oölite (×78) taken from the mud flat west of Ogden, 1,000 feet inland. Very irregular in shape.

FIG. 4.—Oölite (×125) taken from mud flat on west shore of Stansbury Island, 15 feet inland from shore. The nucleus is opaque, surrounded by a highly crystalline zone. Two zones.

FIG. 5.—Oölite (×110) taken from the mud flat, west shore of Stansbury Island, 100 feet inland from shore. The nucleus is made up of three crystals of quartz. Three zones.

FIG. 6.—Oölite (×95) taken from the botton of the lake at north end of Stansbury Island. One zone.

FIG. 7.—Oölite (×70) taken from the mud flat on west shore of Stansbury Island, 1,500 feet inland from shore. Six zones. The shape of the nucleus persisted through most of the zones.

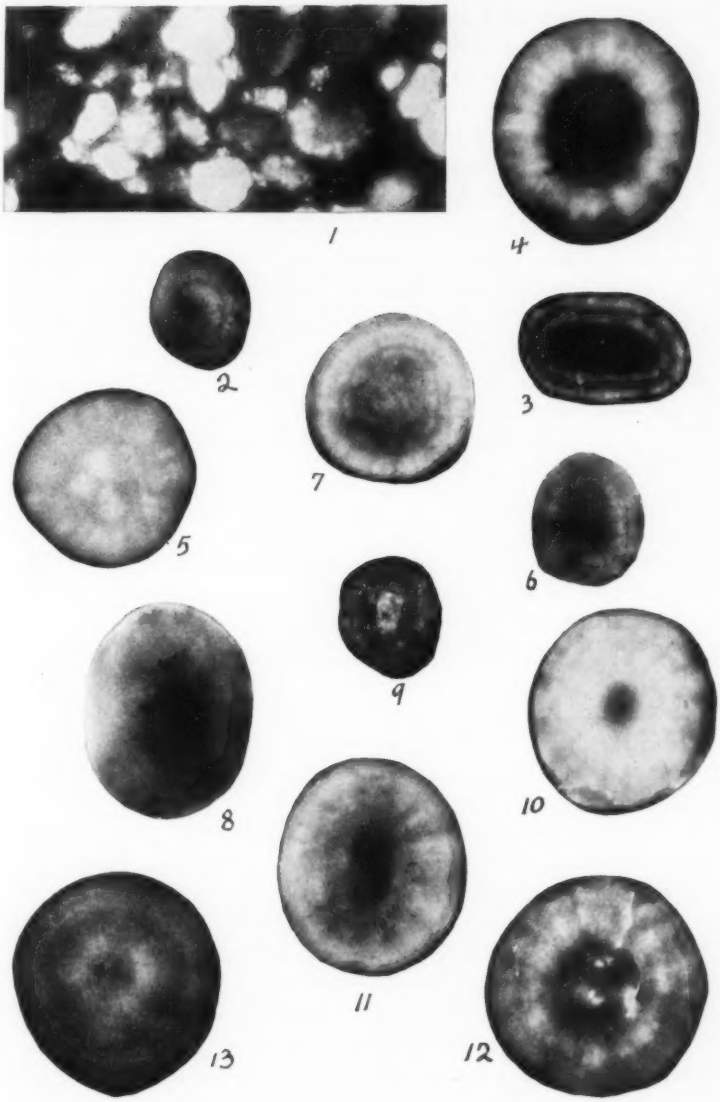
FIG. 10.—Oölite (×85) taken from mud flat on west shore of Stansbury Island, 500 feet inland from shore. Five zones.

FIG. 11.—Oölite (×80) taken from the rocky shore of Antelope Island. Six zones. Note the soot particles on the peripheries of the third and fourth zones.

FIG. 12.—Oölite (×85) taken from the mud flat on west side of Stansbury Island, 2,350 feet inland from shore. Six zones. Several rings within the zones are crossed by the crystals.

FIG. 13.—Oölite (×85) taken from the mud flat near Saltair, 1,000 feet inland from shore. Six zones. Note the nucleus is a group of soot particles mixed with mud.

PLATE I



The process is aided greatly by the difference in temperature between the water in the mud and the capillary water on the oölites. Frequently this difference is as much as 40° F., which is sufficient to cause the calcium carbonate to precipitate about the oölites, forming a very thin lamina of amorphous aragonite. The presence of H_2S does not appear to affect this chemical process.

What chemical action takes place in an oölite during the colder part of the year is problematical; there is, however, a likelihood that the amorphous aragonite is changed to crystalline calcite. At any rate, secondary alteration occurs in many oölites, because it is common for a single crystal to cross two or more zones (see Pl. II, Fig. 8). Murray and Irvine showed[1] that temperature is the deciding factor between the deposition of calcite and aragonite. This might explain the varied thickness of the laminae in the Great Salt Lake oölites, but it will not explain the radial structure, which is post-formational.

EVIDENCE OF GROWTH

In connection with the study of the Great Salt Lake oölites, collections were made to ascertain if there was a gradation in size from the shore inland. The best assortment of material occurred on the east side of the lake and on the west side of the islands in the lake. A gradation in size was observed in all places where the winds were reasonably constant. The samples for detailed study were taken from the mud flat on the west shore of Stansbury Island. Here the conditions were most favorable for natural assortment and deposition, and represent in general the condition elsewhere. The analysis is given in Table II.

Table II shows distinct assortment corresponding to the distance traversed inland. The increase in size is reasonably constant up to the inland lagoon, where the percentage of finer oölites increases because of the local influence of shallow water.

The dune sand also shows a great mixture in size, which is due to the fact that as the wind blows landward some of the smaller oölites are carried along more rapidly than the larger forms.

Thin sections made from the different samples in the foregoing revealed an increase in the number of laminae or zones corresponding

[1] J. Murray and R. Irvine, "Experiments," *Proc. Royal Soc. Edinburgh*, Vol. XVII (1890), pp. 79–109.

to increase in size of the oölites as represented in Table II (see Pls. I and II and descriptions). The variability in size of the dune sand shows in the thin sections. The range in the latter is from oölites having two zones to those having as many as twelve (see Pl. II, Figs. 7, 8, 9, and 10).

The presence of many specks of black carbon included on the peripheries of many zones also offers evidence that oölites grow on land. One of the largest oölites shown on the photomicrographic

TABLE II

ANALYSIS OF SAMPLES OF OÖLITIC SAND OBTAINED FROM THE WEST SHORE OF STANSBURY ISLAND: THE FIGURES ARE GIVEN IN PERCENTAGES OF SAND WHICH PASSED THROUGH THE STANDARD MESH SCREENS

DESCRIPTION OF SAMPLE ANALYZED	SIZE OF SCREEN						
	20	40	50	60	70	80	100
Taken from the water just off shore	.....	.....	.....	.....	77	.....	23
Taken from the bank at water's edge	.....	.....	.....	45	.....	38	17
Taken from mud flat 15 feet inland	9	14	36	29	12	.....	.....
Taken from mud flat 100 feet inland	16	34	24	18	7	.....	1
Taken from mud flat 500 feet inland	27	30	19	14	4	4	2
Taken from mud flat 1,000 feet inland	32	32	20	9	5	2	.....
Taken from mud flat 1,500 feet inland	38	31	16	7	3	3	2
Taken from mud flat 2,350 feet inland at the margin of a lagoon	4	2	8	5	6	24	41
Taken from dune	9	18	21	17	15	14	6

plates was obtained from the lee side of a small obstruction, on the mud flat not more than 400 feet from shore, near Saltair (see Pl. II, Fig. 5). Apparently, it had been rolled about through several periods of changing winds and had not been deposited into a dune. Although this oölite does not possess as many laminae as some, it shows the cross-section of its pitted surface on the third and fourth zones. These pits contain soot which had been picked up from the surface of the mud flat. Discoloration by soot is a common characteristic of the oölites obtained nearest Salt Lake City. The explanation is simple. The soot merely settles from the smoky atmosphere of Great Salt Lake valley during the winter months where it becomes a part of the surface of the mud flats. In the spring when the oölites are driven along by the wind, they become coated with carbon. This process can occur only on land.

EXPLANATION OF PLATE II

FIG. 1.—Oölite (×95) taken from the mud flat on west shore of Stansbury Island, 1,000 feet inland from shore. Six zones. Note the extremely large nucleus with a few soot particles scattered around it. Note also the different thicknesses of the zones.

FIG. 2.—Oölite (×95) taken from the mud flat on west shore of Stansbury Island, 2,350 feet inland from shore. Seven zones. Note the very large, well-defined crystals in the different zones.

FIG. 3.—Oölite (×75) taken from the mud flat on the west shore of Stansbury Island, 100 feet inland from shore. Four zones.

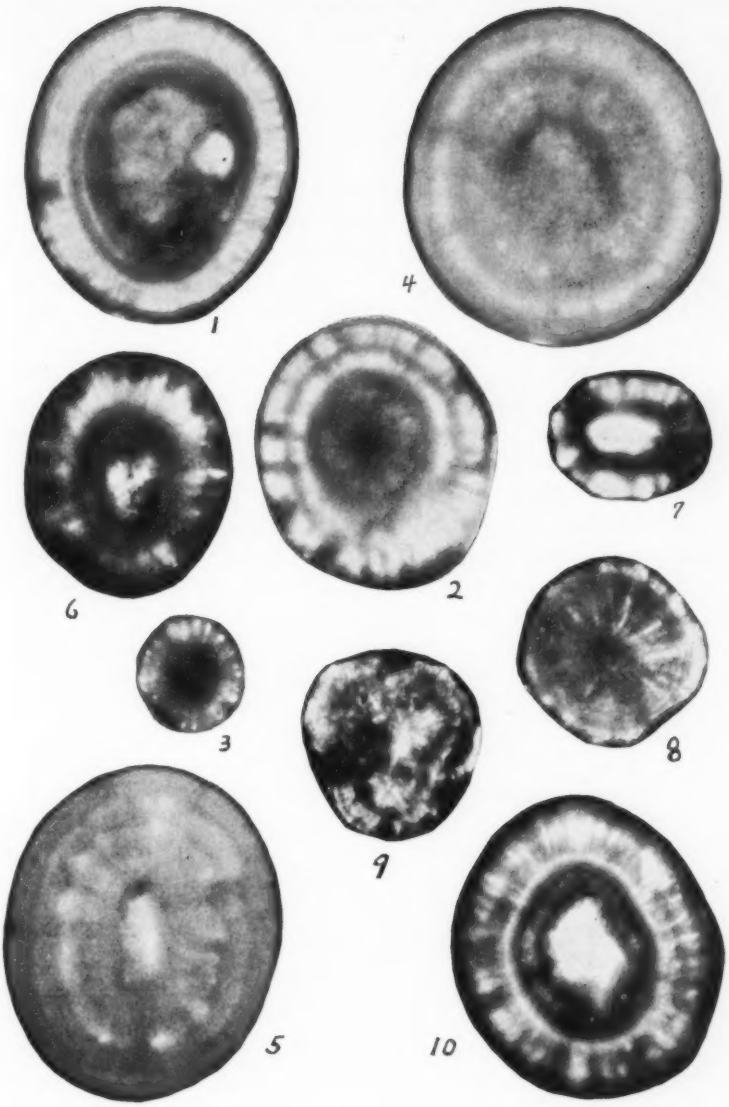
FIG. 4.—Oölite (×90) taken from the mud flat on west shore of Stansbury Island, 1,800 feet inland from shore. Nine zones.

FIG. 5.—Oölite (×78) taken from the lee side of an obstruction on the mud flat near Saltair. Note the pitted periphery of the fourth zone filled with soot particles.

FIG. 6.—Oölite (×82) taken from the mud flat on west shore of Stansbury Island, 1,000 feet inland from shore. Five zones.

FIGS. 7, 8, 9, AND 10.—Oölites (×75) taken from a dune on the west shore of Stansbury Island. These show a variation in size and number of zones.

PLATE II



CONCLUSIONS

The conclusions reached from the study of the Great Salt Lake oölites may be summarized under the following heads:

1. Oölites are formed near the margin of the lake where aëration of the water is greatest.

2. The process is physico-chemical in nature.

3. Growth of the oölites takes place on the mud flats.

4. Oölite-zoning is due to the seasonal changes of the area. (*a*) Period of growth corresponds to the summer season. (*b*) Oölites undergo physical changes and chemical alteration during the winter months.

5. The oölitic sands are wind controlled. The most extensive deposits are on the east shore of the lake or the west shore of the islands, owing to the influence of the prevailing westerly winds.

6. The accumulation of soot on the exterior of the crystalline zones in the oölites is proof of growth on land. It also indicates that each layer of soot was gathered after the rainy season. This suggests that the zones may be annual.

EROSION SURFACES IN IDAHO: DISCUSSION

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Alfred L. Anderson's recent illuminating paper[2] has again brought to attention the often discussed question of the age of remnants of erosion surfaces in Idaho. He refers to the fact that my work in the south-central part of the state has yielded conclusions that appear to be markedly different from those he has arrived at farther north, and suggests various ways to account for the seeming discrepancies. As my conclusions have been published only in abstract form[3] and most of the evidence on which they are based is contained in reports in preparation for publication by the United States Geological Survey, which will not be available for some time, it seems well to take this opportunity to state my viewpoint on these matters.

Anderson presents evidence for the existence in northern Idaho of a summit peneplain (late Cretaceous), a subsummit peneplain (early Tertiary), and a third surface, represented by remnants of old-age valleys, which he says are older than the Latah formation (Miocene). The last statement is based on Kirkham's assignment of sediments near St. Maries to the Latah,[4] but inasmuch as Kirkham found no fossils or other direct evidence as to age, this assignment carries little weight. Kirkham and Johnson[5] have recently pro-

[1] Published by permission of the director of the United States Geological Survey.

[2] A. L. Anderson, "Cretaceous and Tertiary Planation in Northern Idaho," *Jour. Geol.*, Vol. XXXVII (1929), pp. 747–64.

[3] C. P. Ross, "Salient Features of the Geology of South-Central Idaho" (abstract), *Washington Acad. Sci. Jour.*, Vol. XVIII (1928), pp. 267–68.

[4] V. R. D. Kirkham, "Ground Water for Municipal Supply at St. Maries, Idaho," *Idaho Bur. Mines and Geol. Pamph. 17* (November, 1926), p. 4.

[5] V. R. D. Kirkham and M. M. Johnson, "The Latah Formation in Idaho," *Jour. Geol.*, Vol. XXXVII (1929), pp. 483–504.

posed an extension of the definition of the Latah formation originally given by Pardee and Bryan, but the localities they mention in the region covered by Anderson have not yielded identifiable fossils, and therefore their redefinition of the Latah has little bearing on the present discussion. It has been shown[1] that basalt distinctly younger than the Columbia River basalt exists in this general region. Regarding the age of the younger basalt, Pardee and Bryan say: "So far as positive evidence goes the flows in the valleys may be as young as early Pleistocene. Probably, however, they belong to the same period of volcanism that produced Columbia Plateau and other igneous features in the Northwest that are thought generally to belong within the Tertiary." The assumption that the flows with intercalated sediments near St. Maries belong to this younger group seems quite as plausible as the assignment of Latah age made by Kirkham, especially as Pardee and Bryan[2] make the suggestion that similar flows with associated sediments in the valley of the Coeur d'Alene River in northern Idaho, described by Hershey,[3] are to be correlated with "the last or highest flow of Columbia Plateau." In addition it may be noted that Berry's recent work[4] indicates that the Latah itself may be Upper Miocene, rather than Middle or Lower as originally believed.[5] Hence, so far as positive evidence goes, the time limits within which the dates of Anderson's three surfaces must fall are wider, and his last surface may perhaps be much younger than he inferred.

Many will be inclined to doubt the preservation of a peneplain as old as Cretaceous in a part of the northern Rocky Mountain province which has been exposed to erosion since Cretaceous time. The supposed Cretaceous peneplain in British Columbia, which furnishes Anderson's principal basis for the date of his summit pene-

[1] J. T. Pardee and Kirk Bryan, "Geology of the Latah Formation in Relation to the Lavas of the Columbia Plateau near Spokane, Washington," *U.S. Geol. Surv. Prof. Paper 140* (1926), pp. 13–15.

[2] J. T. Pardee and Kirk Bryan, *ibid.*, p. 6.

[3] O. H. Hershey, *Geol. Soc. Amer. Bull.*, Vol. XXIII (1912), p. 529.

[4] E. W. Berry, "A Revision of the Flora of the Latah Formation," *U.S. Geol. Surv. Prof. Paper 154* (1929), pp. 234–35.

[5] F. H. Knowlton, "Flora of the Latah Formation of Spokane, Washington, and Coeur d'Alene, Idaho," *U.S. Geol. Surv. Prof. Paper 140* (1920), p. 23.

plain in northern Idaho, was in large part covered and protected by Eocene and later Tertiary deposits, both sedimentary and volcanic. Even this surface was much dissected at the time the Eocene sediments were laid down.[1]

In south-central Idaho there are widespread and thick volcanic deposits which probably at one time covered the greater part of the mountain region south of the forty-fifth parallel and extended somewhat north of that limit. More detailed mapping has shown the inadequacy of Umpleby's conception[2] that these deposits were restricted to deep canyons cut in a peneplain which he regarded as of Eocene age. Hence Anderson's suggestion[3] that his third surface might be correlated with these filled valleys is inapplicable. The clastic deposits which Umpleby regarded as lake beds filling some of these valleys have been found to be tuffaceous beds forming an integral part of the volcanic series. Some of the tuffaceous material may have been deposited in lakes, but the greater part has no genetic relation to lakes. This is the material which contains the fossil plants mentioned in the foregoing.

As Anderson indicates, my work in south-central Idaho has shown that the only recognizable remnants of an old peneplain preserved in the existing topography truncate and hence are younger than the volcanic strata just mentioned. These strata have generally been supposed to be of Miocene age, and fossil flora found during the course of my work in the upper part of the formation appear to confirm this assignment. Study of the collections by E. W. Berry and R. W. Brown leads them to the belief that the plants represented are

[1] C. W. Drysdale, "Geology of Franklin Mining Camp, British Columbia," *Can. Geol. Surv. Mem. 56* (1915), pp. 18–44; S. J. Schofield, "Geology of Cranbrook Map Area, British Columbia," *Can. Geol. Surv. Mem. 76* (1915), pp. 102, 160–65; C. W. Drysdale, "Geology and Ore Deposits of Rossland, British Columbia," *Geol. Surv. Can. Mem. 77* (1915), pp. 184–88; W. L. Uglow, "Cretaceous Age and Early Eocene Uplift of a Peneplain in Southern British Columbia," *Geol. Soc. Amer. Bull.*, Vol. XXXIV (1923), pp. 561–72.

[2] J. B. Umpleby, "An Old Erosion Surface in Idaho; Its Age and Value as a Datum Plane," *Jour. Geol.*, Vol. XX (1912), pp. 139–47; "Geology and Ore Deposits of Lemhi County, Idaho," *U.S. Geol. Surv. Bull. 528* (1913), pp. 35–40; "Ore Deposits in the Sawtooth Quadrangle, Blaine and Custer Counties, Idaho," *U.S. Geol. Surv. Bull. 580* (1914), pp. 228–29.

[3] A. L. Anderson, *op. cit.*, p. 755.

such as may well be of Miocene age. However, much of the material is poor, and the determinations are consequently doubtful. The forms determined by the two paleobotanists are *Juglans egregia*(?) Lesquereux, *Typha*(?), *Taxodium dubium* (Sternberg) Heer, *Sequoia affinis* Lesquereux, *Ulmus* sp., and *Sapindus*(?) *coloradensis* Cockerell. All these forms are known to have a considerable range within the Tertiary. Other plants represented in the collections were too poorly preserved for even generic identification. It appears that there is nothing in the collections of plant remains so far obtained to preclude the assumption that the strata from which they came belong to an epoch in the Tertiary distinctly earlier than the Miocene. The results of my recent work indicate that the Tertiary volcanic strata in south-central Idaho rest on an erosion surface which is in part nearly level, but in numerous localities had marked relief before it was covered. The volcanic rocks have been folded and faulted, intruded by granite, eroded almost or quite to peneplanation, again uplifted with renewed faulting, and intricately and deeply carved during a series of incomplete erosion cycles. This complex history, which can only be stated here in bare outline, surely suggests a greater age than that of the relatively undeformed Latah formation and Columbia River basalt. One of the striking features of the topography of south-central Idaho is the presence of large basins whose subdued topography is in sharp contrast to the mountains that border them. They may, as has been suggested,[1] owe their form in part to structural control, but as none of the postulated faults have yet been found, this suggestion is problematical. They probably record one stage in the relatively recent erosional history of the region.

The data above outlined lead to the following suggestions: (1) If the evidence for a summit peneplain in northern Idaho is sufficient to withstand the objections of Daly[2] and others to the interpretation

[1] Waldemar Lindgren, "The Mining Districts of the Idaho Basin and the Boise Ridge, Idaho," *U.S. Geol. Surv. 18th Ann. Rept.* (1898), Part 3, pp. 696–98; E. L. Jones, "Lode Mining in the Quartzburg and Grimes Pass Porphyry Belt, Boise Basin, Idaho," *U.S. Geol. Surv. Bull. 640* (1917), pp. 89–90; J. B. Umpleby and D. C. Livingston, "A Reconnaissance in South-Central Idaho," *Idaho Bur. Mines and Geol. Bull. 3* (1920), p. 13; S. M. Ballard, "Geology and Ore Deposits of the Rocky Bar Quadrangle," *Idaho Bur. Mines and Geol. Pamph. 26* (1928), p. 30.

[2] R. A. Daly, "The Accordance of Summit Areas among Alpine Mountains; the Fact and Its Significance," *Jour. Geol.*, Vol. XIII (1905), pp. 105–25.

of a peneplain merely on the basis of approximate accord in summit levels, it should be correlated with the surface below the Tertiary volcanic rocks in south-central Idaho. That surface in the dissected form it had when it was buried is of early Tertiary age, presumably Eocene or Oligocene, but the relatively level portions of it may be remnants of a pre-Tertiary peneplain. (2) The subsummit peneplain of Anderson may well correspond to the summit peneplain or partial peneplain of south-central Idaho. It would then be of probable late Tertiary age, although perhaps not so young as Pliocene, as suggested in my preliminary paper. (3) The youngest surface discussed by Anderson would probably correspond to the numerous broad basins of south-central Idaho, of which the Boise Basin is an outstanding example, and hence would probably be Pliocene or Pleistocene. Thus it appears that, by allowing some latitude in present tentative age assignments, Anderson's observations in northern Idaho and mine in the south-central part of the state can be correlated without calling upon "two generations of mountains," as Anderson feels is necessary.

DISCUSSION BY GEORGE R. MANSFIELD[1]

A. L. Anderson's paper on Cretaceous and Tertiary planation in northern Idaho and C. P. Ross's discussion thereof permit a more satisfactory interpretation of the physiographic history of the state than has hitherto been possible. If the ages of the erosion surfaces described by Anderson are adjusted to the views of Ross, a tentative correlation of these surfaces with those described for southeastern Idaho[2] may be made.

The Salt Lake formation of southeastern Idaho, which is at present regarded as Pliocene (?) but may be older, contains considerable volcanic material and is thus similar in some respects at least to Ross's volcanic deposits of central Idaho, though such plant remains as have been recovered from the Salt Lake formation have not thus far been identifiable. This formation lies in valleys and covers low hills that were formed by the partial dissection of the

[1] Published by permission of the director, United States Geological Survey.

[2] G. R. Mansfield, "Geography, Geology, and Mineral Resources of Southeastern Idaho," *U.S. Geol. Surv. Prof. Paper 152* (1927), pp. 15, 16, 203, 204.

oldest recognized erosion surface of southeastern Idaho—the Snowdrift peneplain. Apparently this peneplain has not been well enough preserved in central Idaho to be recognized, but in the light of Ross's discussion it may very well correspond with Anderson's summit peneplain in northern Idaho.

Anderson's subsummit peneplain of northern Idaho, which Ross thinks may represent the summit peneplain or partial peneplain of central Idaho, may then be recognized in southeastern Idaho as the Gannett erosion surface of supposed late Pliocene or early Quaternary age.

The youngest surface discussed by Anderson, the one responsible for the broad basins that antedate the present canyons, corresponds very well with the Dry Fork surface of southeastern Idaho, which has produced basins that occupy similar positions with respect to existing canyons there. The intervening Elk Valley erosion surface of southeastern Idaho is more local and less well represented there than are the others and may be unrecognizable or absent in central and northern Idaho.

The latest or Blackfoot erosion cycle of southeastern Idaho would then correspond with the present canyons of the central and northern regions.

DISCUSSION BY ALFRED L. ANDERSON

Clyde P. Ross has suggested that, by allowing some latitude in present tentative age assignments, a correlation might be made between the erosion surfaces in the south-central part of Idaho which he has studied and those in the northern part of the state.[1] He suggests that the summit peneplain which I have described might be correlated with the surface below the Tertiary volcanics in south-central Idaho; the subsummit peneplain with the summit peneplain, or partial peneplain, across the Tertiary volcanics; and the youngest surface with broad basins of probable Pliocene or Pleistocene age such as occur at Boise Basin and at many other places.

Such a correlation as Ross suggests would meet with a few ob-

[1] A. L. Anderson, "Cretaceous and Tertiary Planation in Northern Idaho," *Jour. Geol.*, Vol. XXXVII (1929), pp. 747–64.

stacles, particularly in relating the summit peneplain in south-central Idaho to the subsummit peneplain in the north, and referring the lowest surface in the north to the Pliocene or Pleistocene. Had he correlated the summit peneplain in south-central Idaho with the lowest or youngest surface in the north, his suggestions would have harmonized better with the field relations now known to exist in the northern part of the state.

The youngest surface in the north is certainly not cut across the lava of the Columbia Plateau and, consequently, much doubt exists as to whether this surface can be as young as Pliocene, or Pleistocene as Ross suggests. The ridges of this dissected surface stand about 800 feet above the basalt plateau in the St. Joe drainage basin in the Cœur d'Alene Mountains. The same surface is extensively developed along the western margin of the Clearwater Mountains where the level of the summit ridges is approximately 1,000 feet higher than the basalt surface. Of course, should the Columbia Plateau prove to be young as Pliocene or Pleistocene, the erosion surface might be nearly as young, and more latitude might be permitted for age assignments. It is therefore desirable to know the age of the surface flows of Columbia River basalt in order to determine more precisely the age of the erosion surfaces that exist in northern Idaho.

Berry's recent work[1] indicates that the Latah formation is probably Upper Miocene, from which it might be assumed that the associated Columbia basalt is also mainly Upper Miocene. This age, for the surface flows, might be questioned, inasmuch as basalt distinctly younger than the Columbia River basalt has been shown[2] to exist in the general region. But until the precise age of these younger flows is determined and until it is definitely proved that the young basalt actually forms the plateau surface along the margin of the Cœur d'Alene and Clearwater mountains, there is no justification for assigning a Pleistocene age to the Columbia Plateau which would be necessary if Ross's suggested correlation of the lowest surface is

[1] E. W. Berry, "A Revision of the Flora of the Latah Formation," *U.S. Geol. Surv. Professional Paper No. 154* (1929), pp. 234–35.

[2] J. T. Pardee and Kirk Bryan, "Geology of the Latah Formation in Relation to the Lavas of the Columbia Plateau near Spokane, Washington," *U.S. Geol. Surv. Professional Paper No. 140* (1926), pp. 13–15.

accepted. On the other hand, the dating of the erosion surface from the top flows of basalt might not be so important as one might first suppose, inasmuch as the Latah formation bears the same relation to the dissected erosion surface as does the overlying basalt. The lowest surface on the north must therefore have been carved not later than the deposition of the Latah sediments, or not later than the Upper Miocene.

Whether the lowest erosion surface in northern Idaho can be correlated with the summit peneplain in south-central Idaho depends also on the age of the Tertiary volcanics across which the summit peneplain is cut. Should the Tertiary volcanic series prove to be older than the Columbia basalt, the suggested correlation might well apply.

The volcanic strata in south-central Idaho have generally been supposed to be of Miocene age, but Ross indicates that nothing in the collections of plant remains so far made precludes the assumption that the plant-bearing strata are distinctly earlier than Miocene. Confirmation of this assumption might be verified if the Tertiary volcanics and the Columbia basalt were to be found in contact.

My field work[1] in the Clearwater Mountains and adjacent Columbia Plateau during the summer of 1929 perhaps offers evidence of an age difference between the two series of extrusives. An exposure of volcanic strata greatly resembling the Tertiary volcanics in south-central Idaho was found unconformably below the Columbia basalt and, indeed, projecting above it. The exposure comprises Kamiah buttes, in Idaho County, lying about 15 miles from the edge of the Clearwater Mountains and rising about 1,000 feet above the plateau surface from a base covering 22 square miles. The volcanics rest on the eroded surface of an outlier of the Idaho batholith, disclosed in a canyon carved across the south side of the buttes. The volcanic strata have the composition of andesite and latite, resembling the volcanics of south-central Idaho studied in the Lava Creek district.[2] The flows have not been much disturbed and are es-

[1] A. L. Anderson, "The Geology and Mineral Resources of the Region about Orofino, Idaho," *Idaho Bureau of Mines and Geology Pamphlet No. 34* (1930).

[2] Anderson, "Geology and Ore Deposits of the Lava Creek District, Idaho," *Idaho Bureau of Mines and Geology Pamphlet No. 32* (1929), pp. 13–19.

sentially flat lying. Their much-eroded slopes are in marked contrast with the flat surface of the plateau.

A series of extrusives of intermediate composition, older than the Columbia basalt, is therefore demonstrated, but the precise difference in age has not been determined. The rocks of Kamiah buttes probably belong to the same period of volcanism as those of south-central Idaho, although from present knowledge this correlation must be based on lithology alone. Probably the interval between the two events is sufficient to allow partial peneplanation of the Tertiary volcanics before the extrusion of the latest flows of Columbia basalt. An assignment of the Tertiary volcanics to the lower and even to the middle of the Miocene would probably allow sufficient time for this bit of erosional history.

Thus, if Ross's correlation be changed so as to relate the summit peneplain of south-central Idaho to the youngest or lowest erosion surface in the northern part of the state, several serious objections would be overcome. This would place the subsummit surface, as well as the summit, if any of it remains in south-central Idaho, beneath the Tertiary volcanics, and the youngest erosion surface (pre-Columbia Plateau) across the Tertiary volcanics. The summit and subsummit surfaces may not be so old as I suggested, but perhaps ample time existed for their development within the Tertiary without going so far back as the Cretaceous as I felt compelled to do.

Correlation of the surfaces in the south-central with those in the northern part of the state is still a matter of speculation and must remain so until a rather detailed reconnaissance is made of the intervening area, i.e., in the region comprising the southern Clearwater Mountains and the northern Salmon River Mountains. More attention should also be given to the relation of the youngest flows of basalt to the Latah formation, as well as to a more precise dating of the Tertiary volcanics series and its relation to the Columbia basalt.

OLD EROSION SURFACES IN SOUTHWESTERN IDAHO[1]

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Lansing, Michigan

ABSTRACT

In southwestern Idaho there is strong evidence pointing to a pre-Columbia River basalt peneplain which may be equivalent to Anderson's Eocene surface. A post-basalt erosion surface of late Miocene or early Pliocene age, which at some places was not a true peneplain, is also indicated. Two Pleistocene erosion surfaces are well preserved in the Snake River Basin. All four erosion surfaces have been considerably warped from their original positions. Evidence in northern Idaho confirms Anderson's previous interpretation of pre-Columbia River lava erosion surfaces.

INTRODUCTION

The area studied is that part of southwestern Idaho lying between meridians 116°20′ and 117°15′ and parallels 43°22′ and 44°30′. It comprises about ninety townships. The writer gratefully acknowledges the helpful suggestions of Dr. Edson S. Bastin, Dr. J Harlen Bretz, and Dr. Rollin T. Chamberlin, all of the Department of Geology, University of Chicago, and the determination of flora by Dr. Erling Dorf, Princeton University.

PRE-COLUMBIA RIVER BASALT EROSION SURFACE

The oldest erosion surface exposed within the area is believed to be present on the Jurassic (?) granite exposures in the region. Strong evidence exists that the Columbia River basalt and the rhyolite were poured out on a relatively flat granite surface. The Columbia River basalt is believed to be Middle or Upper Miocene, and contains the interbedded Payette lake beds. The rhyolite is of Upper Miocene or Lower Pliocene age, being underlain by sediments containing Miocene flora and overlain by the Idaho lake beds yielding Pliocene flora.

Where the rhyolite overlies the granite east of Squaw Creek in the Squaw Creek quadrangle, the contact is notably flat for a distance of several miles (see Fig. 1). The area of granitic rocks in Boise quadrangle, called Boise Ridge, is capped in various places

[1] This work was done in co-operation with the Idaho Bureau of Mines and Geology, and the results are published with the permission of the secretary.

(including the highest point at 7,200 feet) by small isolated remnants of Columbia River basalt. These basalt remnants, at all places observed, dip concordantly with the surface of the granitic rock. The basin lake beds (Pliocene) overlying the basalt at the foot of the ridge also dip at angles and directions similar to the slope of the granite surface. This relation suggests that these basalts were poured out on a nearly level granitic plain that has been tilted since the Pliocene.

FIG. 1.—Black line indicates contact of rhyolite series on peneplain surface of Jurassic granite in Squaw Creek quadrangle.

Lindgren[1] accounted for the different elevations of these flows by postulating local vents for each of the isolated exposures. Their inclination he believed to represent original flow slopes, and thus explained their accordance of dip with that of the slope of the granite on which they lie. The writer's study did not reveal any local sources for erosional remnants; and he concludes, partly because of other evidence found elsewhere in the area, that the Columbia River basalt and the old erosion surface on which it rests were originally nearly horizontal and have been warped or tilted into their present

[1] Waldemar Lindgren, "Description of the Boise Quadrangle," *U.S. Geol. Surv., Geol. Atlas, Folio 45* (1898).

position. The demonstrable evidence leading to this conclusion is obtained by plotting many surface profiles across Boise Ridge. Elevations are obtained from the topographic sheet, and outcrop elevations of the basalt lake beds overlying the granite are taken from the geologic sheet of Boise Folio. Sections constructed from these data, and with actual dip-angles of basalt flows and bedding

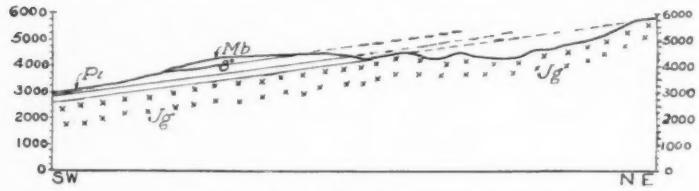


FIG. 2.—Section drawn to scale showing Columbia River basalt (*Mb*) contact on peneplain surface of Jurassic granite (*Jg*) in Boise quadrangle. Note that angle of dip in basalt and overlying Idaho formation (*Pi*) is similar to flats on the granite.

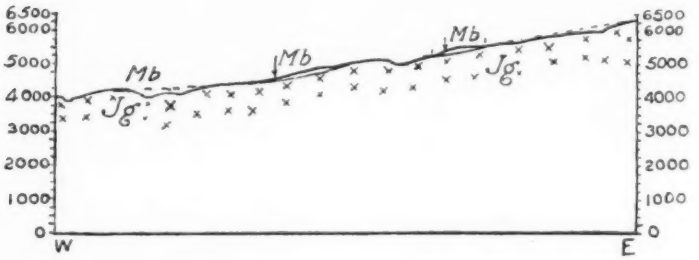


FIG. 3.—Section with equal scales showing basal flow of Columbia River basalt (*Mb*) lying on peneplain on Jurassic granite (*Jg*) in Boise quadrangle. Dotted line indicates warped peneplain surface.

planes measured in the field, produce results represented by Figures 2 and 3. Several lines of evidence shown by these profile sections are:

1. The slope of the pre-Columbia River basalt surface of the granite. This can be determined by connecting, in profiles drawn to scale, the present levels of the base of the basalt on the granite. Where this is done, as in Figure 3, it is clear that the granite surface

is nearly a plane with a maximum dip of about eight degrees to the southwest.

2. No granite prominences rise above a surface so determined.

3. The slope of land surfaces underlain by granite above corresponds closely with the slope represented by the base of the basalt. This indicates that, in such areas, the granite has been only slightly eroded since the basalt cover was removed by erosion.

4. Actual dips within the basalt and the overlying sediments, when plotted to scale, are similar in angle and direction to the slopes obtained by plotting the base of the basalt as in paragraph 1.

Sections measured in the same direction showed marked uniformity in slope, although the basalts and overlying sediments vary slightly from place to place in their direction of dip. The erosion surface on the granite is probably not a plane, but rather a surface in which warping has produced gentle undulations. Thus at some places the surface appears to dip due west and at others several miles removed, it apparently dips W. 15° S.

This old erosion surface illustrated by the line between the granite and the basalt in Figures 2 and 3 appears to represent a peneplain of pre-Columbia River basalt age. The writer, for several years, has been aware of at least two pre-Columbia River basalt erosion surfaces appearing in other parts of Idaho; and Anderson,[1] in the most recent paper on Idaho planation, supports this view by presenting evidence for three erosion levels in northern Idaho, all of which are of pre-basalt age. He also supports the convictions long held by the writer, that the physiographic history of different parts of Idaho has not been the same, and that all stages of the history of the erosion surfaces are not everywhere represented. Anderson classes his pre-basalt peneplains as late Cretaceous, late Eocene, and early Miocene. Inasmuch as the granite on Boise Ridge, a part of the Idaho batholith, is now believed to be of Jurassic age, the erosion surface recognized here may be any one of Anderson's three stages. Because of its low relief and widespread occurrence, the surface in this area is not believed to be the early Miocene surface of northern Idaho, which is represented there as an irregular surface found only

[1] Alfred L. Anderson, "Cretaceous and Tertiary Planation in Northern Idaho," *Jour. Geol.*, Vol. XXXVII (1929), pp. 747–64.

in wide valleys. The writer is inclined to assign this peneplain to the late Eocene and to correlate it with Anderson's middle stage. It is very likely one of two surfaces recognized by Umpleby,[1] Blackwelder,[2] Atwood,[3] Rich,[4] Livingston,[5] Lindgren,[6] Mansfield,[7] and Buwalda,[8] and described by them in the spirited discussion of "The Old Erosion Surface in Idaho." This debate probably arose because the participants all believed that only one peneplain existed. So far as is known, the writer[9] preceded other observers in recognizing a post-Miocene surface as well as a pre-Miocene surface in south-central Idaho.

POST-COLUMBIA RIVER BASALT EROSION SURFACE

Evidence for a post-Columbia River basalt, pre-Idaho formation (Pliocene) erosion surface, is also present in the area. From the brink of the Snake River Canyon, in the Huntington quadrangle, one can see a notably level surface extending eastward across the folded basalt (see Fig. 4).

That the lowest basin beds of the Idaho formation were deposited on the Columbia River basalt when it was practically horizontal is attested by the following lines of evidence:

[1] Joseph Umpleby, "An Old Erosion Surface in Idaho," *ibid.*, Vol. XX (1912), pp. 139–47; "The Old Erosion Surface in Idaho," *ibid.*, Vol. XXI (1913), pp. 224–31.

[2] Eliot Blackwelder, "The Old Erosion Surface in Idaho," *ibid.*, Vol. XX (1912), pp. 410–14; "Physiographic Conditions and Copper Enrichment," *Econ. Geol.*, Vol. XII (1917), pp. 541–45.

[3] W. W. Atwood, "The Physiographic Conditions at Butte, Montana, and Bingham Canyon, Utah, When Copper Ores in These Districts Were Enriched," *Econ. Geol.*, Vol. XI (1916), pp. 697–740; "Physiographic Conditions and Copper Enrichment," *ibid.*, Vol. XII (1917), pp. 545–47.

[4] J. L. Rich, "An Old Erosion Surface in Idaho; Is It Eocene?" *ibid.*, Vol. XIII (1918), pp. 120–36.

[5] D. C. Livingston, "The Idaho Peneplain (Discussion)," *ibid.*, Vol. XIII (1918), pp. 488–92.

[6] Waldemar Lindgren, "The Idaho Peneplain (Discussion)," *ibid.*, Vol. XIII (1918), pp. 486–88.

[7] G. R. Mansfield, "Tertiary Planation in Idaho," *Jour. Geol.*, Vol. XXXII (1924).

[8] John P. Buwalda, "The Age of the Payette Formation and the Old Erosion Surface in Idaho," *Sci.*, Vol. LX (1924), pp. 572–73.

[9] Virgil R. D. Kirkham, "A Geologic Reconnaissance of Clark and Jefferson and Parts of Butte, Custer, Fremont, Lenhi and Madison Counties, Idaho," *Idaho Bur. of Mines and Geol. Pamph. 19* (January, 1927).

1. The high dips observed in places in the basin beds of the Idaho formation, and the correspondence in dip between the basal lake beds (Idaho formation, Pliocene) and the underlying basalt, indicate that the basalt and basin beds were folded concurrently and to an equal degree.

2. Because the lithologic character of the basin lake beds implies deposition in a nearly horizontal position, it follows that the underlying basalt beds were at that time also nearly horizontal. Practically no basaltic detritus appears at any place in the basin beds.

FIG. 4.—Looking northeast across Snake River at steeply folded Columbia River basalt in Huntington and Weiser quadrangles. Note the striking peneplain-like surface truncating the basalt folds.

3. The basin lake beds were not, however, deposited immediately after the extrusion of the basalt, but after an interval during which notable erosion of the basalt took place, creating marked differences in relatively short distances, in the total thicknesses of the basalt which overlies the Payette formation and which separates the two lake-bed formations.

That considerable erosion had occurred on the surface of the Columbia River basalt and on the rhyolite before the deposition of the basal Idaho lake beds is shown by the notable disconformity between each of them and the basin beds. Speaking of this discon-

formity as seen in Boise and Weiser quadrangles, Lindgren[1] said: "On the west the basalts border against the Payette[2] sandstones and as far as can be seen dip below them forming a very uneven surface upon which the Payette sandstones were deposited."

Buwalda,[3] in describing the relation between the basin beds and the rhyolite on the south sides of the plain said: "The second was found in the same section in strata overlying the rhyolite, probably disconformably."

The best evidence of this disconformity observed by the writer is the notably uneven surface of basalt. Where the lake beds near the basalt or rhyolite have been nearly stripped off by erosion, small protrusions of lava project from the lake-bed surface. Slopes of lava, once covered by lake beds, still retain small uneroded pockets of lake bed material protected in minor depressions.

The irregularity of the northwest-southeast contact line between the basalt and the basin beds in the Boise and Weiser quadrangles and the northwest-southeast contact line between the rhyolite and basin beds in the Manpa and Silver City quadrangle speak strongly for a differentially eroded surface on the lava before deposition of the basin beds. This evidence is particularly convincing in light of the similarity of dip in lavas and the overlying lake beds. At one place west of Snake River on the south edge of the area the Idaho formation rests directly on the Silver City granite; two small peaks of Columbia River basalt protrude through the lake beds indicating its presence beneath. That the Columbia River basalt and the overlying rhyolite were once present and were eroded from most of the granite area before the deposition of the Idaho formation is proven by the basalt cap overlying the granite on Wilson Peak, and the presence of the rhyolite on Reynolds Creek less than a mile to the south.[4] The occurrence of the Idaho formation immediately overlying the granite at the head of Willow Creek in the west part of T. 6 N., R. 1 E., B.M., and in the eastern part of T. 5 N., R. 1 E.,

[1] "Description of the Boise Quadrangle," *U.S. Geol. Surv., Geol. Atlas, Folio 45* (1898), p. 4.

[2] These beds are the Idaho lake beds but were included in Lindgren's Payette formation.

[3] *Op. cit.*, p. 573

[4] See *Silver City Folio 104.*

indicates the same conclusions. That the basalt originally covered the granite in the northern locality and was removed by erosion before the deposition of the Idaho beds is indicated by the presence of basalt to the north and to the south underlying the lake beds and by the small basalt area to the west in Secs. 11 and 12 exposed by erosion of the Idaho beds.

In T. 5 N., R. 1 W., also, the Idaho formation immediately overlies the granite. That basalt intervenes between the two formations at a point some distance west of the contact is suggested by its occurrence in an intervening position a short distance to the south in Secs. 23 and 24. If this is true, the basalt was partly eroded from the granite before the deposition of the Idaho beds, and its remnants are now covered by the overlap of the lake beds. The erosional differences amount, apparently, to several hundred feet.

After the Idaho formation was deposited nearly horizontally on the eroded surface, the entire sequence was folded and warped into its present position and subsequently eroded in the anticlines until the upper Columbia River basalt and the Payette formation had been cut through to the lower Columbia River basalt. This post-Columbia River basalt erosion surface may represent the post-Miocene or Pliocene surface noted by the writer[1] and later by Ross[2] in other parts of southern Idaho. That it was not a true peneplain in this region is indicated by a difference of relief of 600 feet in a distance of 3 miles.

A notably level erosion surface truncates the steeply folded basalt on Indian Head west of Weiser. This level accords with the level surface shown by the photograph (Fig. 4), and may possibly represent a remnant of the post-Miocene erosion surface suggested in the erosion flat in the Huntington area and in the folds in Weiser quadrangle.

UPPER MESA EROSION SURFACE

In the physiographic division of the area known as the Snake River Basin, a Pleistocene base level, or peneplain, is represented by the exceptionally flat surface which truncates the basin beds, and

[1] *Op. cit.*

[2] C. P. Ross, "Salient Features of the Geology of South Central Idaho," Abstract, *Wash. Acad. of Sci.*, Vol. XVIII (1928), pp. 267–68.

on which the Pleistocene Upper Mesa gravels, sand, and silt were deposited. This surface was formed by lateral planation by Payette, Boise, and Snake rivers. Large remnants of this surface are protected by erosional remnants of Upper Mesa gravels. One of these areas, 9 miles wide, lies in Bisuka and Boise quadrangles south of Boise River. This surface lies at elevations ranging from 2,600 to 2,800 feet above sea-level and from 400 to 600 feet above the present rive. Another area, 8 miles wide, lies in Nampa and Weiser quadrangles north of Payette River. The surface here is at elevations ranging from 2,400 to 3,000 feet above sea-level. In each instance the highest elevation is toward the basin edge and the lowest nearest the center of the basin.

South of Boise River in the Nampa quadrangle the surface is protected for a width of $8\frac{1}{2}$ miles by Upper Mesa gravels. The farthest south elevation is 2,800, and the lowest elevation at the northern extremity of this strip is below 2,400 feet. That the surface originally possessed much greater extent than that now covered by the remnants of Upper Mesa is suggested by flat adjacent regions north of Payette River and south of Snake River from which the gravel layer has been eroded. The former existence of the gravel over these flats, which truncate the tilted basin beds (Idaho formation), is indicated by scattered gravel detritus and cobbles identical with the material in the beds of Upper Mesa.

At every place so far studied, this erosion surface upon which the Upper Mesa gravel lies, dips toward the axis of the downwarp which is responsible for the basin, and also in a westerly direction. The well-bedded Upper Mesa formation dips identically with the surface upon which it lies. The inference is that this surface and the Upper Mesa beds have been slightly warped since the original deposition of the Upper Mesa formation.

It is not believed that the surface underlying the Upper Mesa formation extended as a unit over the entire basin area, but rather that the Payette, Boise, and Snake rivers each developed flood plains several miles wide and that in the case of the Payette and Boise rivers the streams now lie closer to the axis of the downwarp than when they first started deposition of the Upper Mesa formation.

No Upper Mesa beds lie south of Payette River, practically none border Boise River on the south, and only three small patches lie to the south of Snake River.

LOWER MESA EROSION SURFACE

A later Pleistocene erosion stage is recorded by another extensive erosion surface at a lower level than that which carries the Upper Mesa formation. This second Pleistocene level also resulted from lateral planation by the Boise, Payette, and Snake rivers. It is almost everywhere preserved by its cover of Lower Mesa gravels. Relatively small amounts of Lower Mesa gravels have been removed by erosion, and the extent of this surface for practical purposes may be said to be the areal outcrop of that formation.

Like the Upper Mesa, the Lower Mesa formation and the surface upon which it lies has been warped but slightly. Total drops greater than 200 feet have not been measured.

DISCUSSION OF RECENT PAPERS

Ross's[1] discussion in the preceding article attacks Anderson's[2] conclusions regarding the age of the erosion surface in the St. Maries area on the assumption that the age of the basalt in that region is not Miocene and may be much younger. He says:

> The assumption that the flows with intercalated sediments near St. Maries belong to the younger group seems quite as plausible as the assignment of Latah age made by Kirkham, especially as Pardee and Bryan make the suggestion that similar flows with associated sediments in the valley of the Cœur d'Alene River in Northern Idaho, described by Hershey, are to be correlated with "the last or highest flow of the Columbia Plateau."

Fortunately it is not necessary to make any assumptions about the presence of the Latah formation in this locality or about the age of the basalts associated with it. The area in question which has been here described in two former publications[3] lies in Sec. 7, T. 45 N., R. 1 W.

[1] Clyde P. Ross, "Old Erosion Surfaces in Idaho (Discussion)," *Jour. Geol.*, Vol. XXXVIII (1930), p. 644.

[2] *Op. cit.*

[3] Virgil R. D. Kirkham, "Ground Water for Municipal Supply at St. Maries, Idaho," *Idaho Bur. of Mines and Geol. Pamphlet 17* (November, 1926); Virgil R. D. Kirkham and M. Melville Johnson, "The Latah Formation in Idaho," *Jour. Geol.*, Vol. XXXVII (1929), pp. 483–504.

Fossils collected and submitted to Dr. Erling Dorf, paleobotanist at Princeton University, since the publication of the former papers have been determined by him as characteristic Latah flora. Among the definite Latah types which were recognized were many good specimens of *Populus heteromorpha* Kn., *Acer bendirei* Lesq., *Lauris grandis* Lesq., *Betula heteromorpha* Kn., *B. Largei* Kn., *Taxodium dubium* (St.) Heer, *Castanea castaneafolius* (Un) Kn., and *Platanus dissecta* Lesq. Dorf,[1] in speaking of this assemblage, said, "In my opinion the assemblage is thus clearly and quite definitely Latah."

These data clearly corroborate Anderson's[2] age assignments and show just how unjustified were Ross's assumptions of "younger" basalt.

Ross[3] says in his discussion, "It has been shown[4] that basalt distinctly younger than the Columbia River basalt exists in this general region." Unfortunately the belief in post-Columbia River basalt for this region is not universally accepted and the chief opponents of this idea include those who have studied the type areas.

As was shown by Kirkham and Johnson,[5] the Latah formation is characteristically interbedded with the Columbia River basalt, rather than underlying it, over a front of about 250 miles. Two series of basalt flows are almost everywhere present; and it is to be granted, of course, that the relatively great thickness of intervening Latah formation represents a period of quiescence of considerable duration between the lava extrusions. That this quiescent period was not a local phenomenon is supported by the existence in southwestern Idaho, over an area equally large, of a similar interbedded lake bed formation[6] (now redefined as the Payette) intervening between an upper and lower series of Columbia River basalt flows. The interbedded formation has already yielded from limited collections twenty-three of the seventy-one species common to the Latah flora. The northernmost outcrop of this formation lies only

[1] Written communication.

[2] *Op. cit.*

[3] *Op. cit.*

[4] By Pardee and Bryan.

[5] *Op. cit.*

[6] This formation will be fully described in a forthcoming paper.

85 miles from the definitely established Latah locality at Whitebird, and may actually represent the same horizon.

That the upper series of flows are no later than Upper Miocene in age is indicated by overlying lake beds which contain Upper Miocene fauna and flora.[1]

The writer is one of those who believe that the Latah formation at Spokane may perhaps also occur as an interbedded formation and that the "valley basalts" are just as likely to represent flows of the upper series of the Columbia River basalt as to represent Pliocene or Pleistocene extrusions. The writer believes that Anderson's[2] analysis of the planation history is corroborated by the field evidence. He is convinced that there is no evidence in northern Idaho for a post-Columbia River basalt peneplain.

[1] This occurrence is in Silver City quadrangle; and full evidence, including fossils determined, will be furnished in a separate paper.

[2] *Op. cit.*

REVIEWS

Geologic Structures. By BAILEY WILLIS and ROBIN WILLIS. Second Edition. New York: McGraw-Hill Book Company. 1929. Pp. xv+518; figs. 152; pls. in appendix 12. Price $4.00.

So much has the second edition of this well-known work grown beyond the limits of the first edition in scope and in treatment, that it comes almost with the freshness of a new book. Splendidly written in the first edition, various sections have been retained unchanged, but in many places where the earlier treatment seemed too brief, the discussion has been rounded out with important additions, or completely reorganized and rewritten in fuller form. To this enlargement has been brought the accumulated experience of world-wide field work and unsurpassed familiarity with the earth's structure. Of great value and wholly new is Section IV, of eighty-seven pages, on major problems of earth dynamics. Here the specialist and advanced student will study the pages eagerly for the senior author's latest points of view and mature judgment on those aspects of our science where the frontier is new and advancing into the unknown. The younger student also will find here that the clarity of text and the masterly handling of evidence and inference carry him steadily on, without undue effort, as far as it is safe to journey at the present time.

Added material in the early pages covers the consolidation of sediments, unconformities, warping, and zones of folded strata wherein three classic types give the student a preliminary picture of rock behavior in folding. The chapter on joints is much expanded, particularly in explanation of how joints develop. From joints the text now continues directly into faults, the old chapter on structures of igneous rocks, which came between joints and faults, having been shifted to a more logical place preceding the structures of metamorphic rocks, where it appears extended and improved. The old chapter on faults has been split into two chapters —descriptions and technical discussions—in which the material has been completely reorganized and amplified, and fault problems pushed farther ahead. The classification as high-angle faults and low-angle faults is logical and undoubtedly sharpens the concepts for the student. Under high-angle faults particular attention is given to the striking cases of the Rhine graben, the Dead Sea trough, and the great African Rift valleys. The rival hypotheses of ramp versus rift are critically compared in a way to afford ideal training in the methods of structural interpretation. Willis

favors the compression and upthrust interpretation of these structures, introducing the new term "ramp" to carry that idea in contrast to rift which connotes tension-tearing.

Low-angle overthrusts are given an excellent treatment which is carried on into a discussion of alternative views of Alpine structure. Bailey Willis' own interpretation of the Préalpes as the result of an older thrust from the northwest later displaced by a younger thrust from the southeast, is brought to the fore as an alternative for consideration. The technical discussion of faults, reserved till the student has been prepared for it by a wealth of descriptive illustration, works admirably. To substitute dip for hade, where that term appears, would, however, seem to the reviewer a possible improvement. Hade, in the opinion of the reviewer, merely adds a possible source of confusion, and the sooner it is allowed to fall into disuse the better. The uniform measurement of angles (slope of surface, dip of strata, dip of dike or fault plane) from the horizontal is simpler and better.

Section I is an exposition of geologic structures as such. A vast amount of interpretation inevitably enters into this, furnishing much of the flesh and blood of a live treatment of the subject. The more systematic analysis of the underlying mechanical principles and processes is, however, reserved for Section II. Originally this comprised a single chapter on the mechanics of rock deformation, but in the new edition the ground is covered in five separate chapters—the problem of rock deformation, the mechanical principles, primary types of deformation, analysis of folding, and analysis of faulting. Worthy of particular notice is the analysis of competent folding contrasted with incompetent folding, much of which is developed from the senior author's classic *Mechanics of Appalachian Structure*.

Section III presents the methods of attack. Fortunately the charming essay on field methods (chapter xiii) remains exactly as it was in the first edition, except for the omission of several pages of fine print which interrupted the continuity. To Robin Willis' chapter on graphic methods and practical problems, there have been added a few pages on maps and structure mapping.

Section IV is the greatest innovation. Recognizing that much of the earth's character and present traits were inherited at its birth and developed during its youth, the senior author has examined critically the theories of earth genesis. In no other phase of geology do the origin and formative stages of the earth enter more basically than in dynamic structural geology. Hence the importance of a clear understanding of the strength and weaknesses of present ideas of earth genesis. Willis carries

along a parallel study of the planetesimal theory versus Jeffrey's modification of Jeans' derivative from it (tidal disruptive theory). He finds the former a more comprehensive analysis, and the latter a special case that lacks some of the factors which would inevitably enter into any natural occurrence. The planetesimal theory, he believes, keeps more closely in touch with the actual conditions as we know them.

The text then proceeds with our present knowledge of the constitution of the earth body, followed by an excellent exposition of the forces which operate on this material. Zones of pressure, strength, and temperature in the outer crust, and dynamic spheres of the globe are the two concluding chapters. Concisely they give the best understanding which we have of the larger dynamic divisions of the globe, with the various lines of evidence upon which the inferences are based, and the underlying principles and theories which have been developed and which contribute in turn to the full picture of the earth as a dynamic body. These chapters merit very careful study and consideration.

The new edition is a much more teachable text than the old. It is easier to use and yet it goes distinctly deeper into the subject. There is, of course, also the advance made possible by the increase of knowledge since the first edition came out in 1923. Smaller pages and a better system of headings improve the appearance of the volume very notably. *Geologic Structures* is a most important contribution to the teaching of dynamic geology and a book of great value to the practical and theoretical geologist alike.

R. T. C.

Tertiary Foraminifera from Humboldt County, California. By JOSEPH A. CUSHMAN AND ROSCOE E. AND KATHERINE C. STEWART. "Transactions of the San Diego Society of Natural History," Vol. VI, No. 2. Pp. 41–94; pls. 1–8; chart. San Diego, February, 1930.

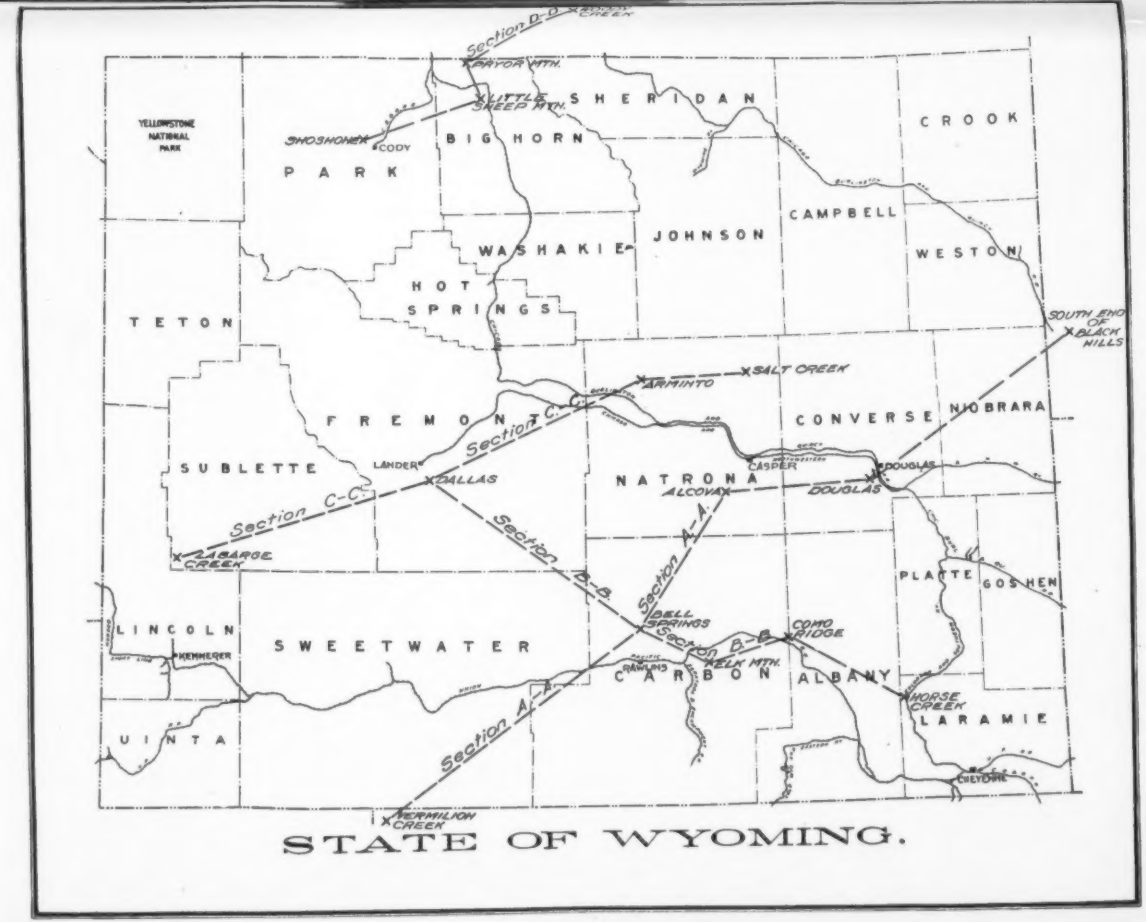
This paper is a preliminary survey of an extensive collection made from forty-eight localities in the Miocene and Pliocene sediments of Humboldt County, California. As here described the fauna consists of 43 genera, 82 species, and 9 varieties. Of these, 5 species are new. The writers are to be congratulated for giving the accurate location of each collecting area, and for indicating the relative abundance of each species at every locality at which it occurs. Lists of those species which are confined (1) to the Miocene and (2) to the Pliocene are also included.

The plates, which were prepared from drawings by Miss Margaret Moore, are very effective.

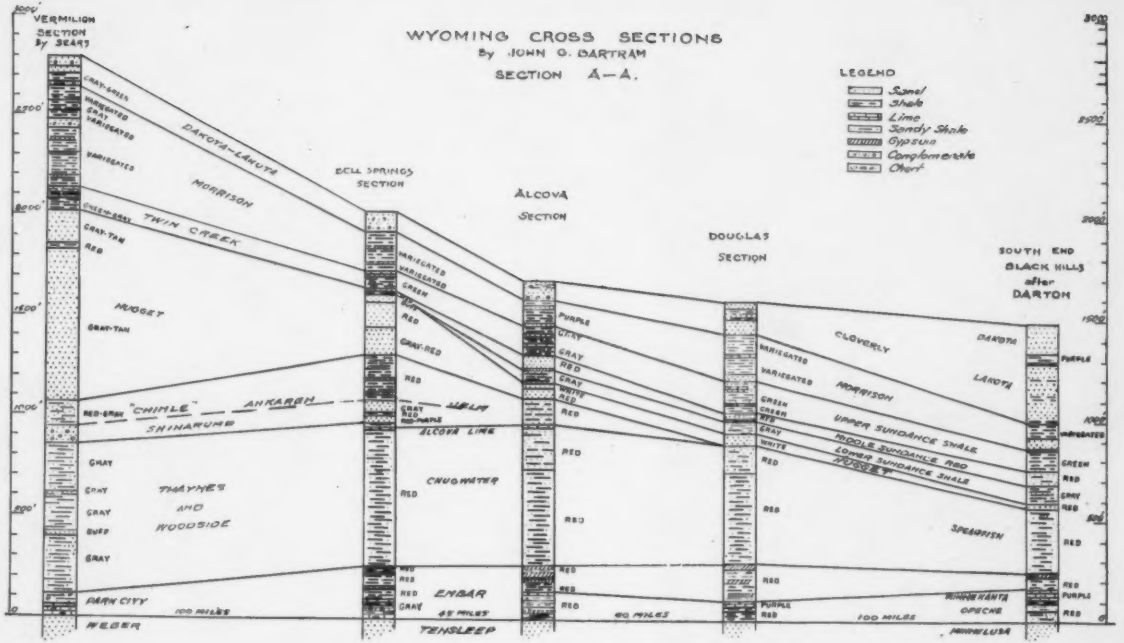
C. C.

ERRATUM

The following illustrations should have appeared in Number 4 of the *Journal of Geology* with the article by John G. Bartram, "Triassic-Jurassic Red Beds of the Rocky Mountain Region." These pages may be torn out and inserted in the proper place in Number 4.



STATE OF WYOMING.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By JOHN G. BARTRAM
SECTION A—A.
LEGEND
Sand
Shale
Lime
Sandy Shale
Gypsum
Conglomerate
Chert
VERMILION SECTION By SEARS
BELL SPRINGS SECTION
ALCOVA SECTION
DOUGLAS SECTION
SOUTH END BLACK HILLS after DARTON
DAKOTA-LAKOTA
MORRISON
TWIN CREEK
NUGGET
"CHIMLE"
ANKAREH
SHINARUMP
THAYNES AND WOODSIDE
PARK CITY
WEBER
ALCOVA LIME
CHUGWATER
EMBAR
TENSLEEP
CLOVERLY
UPPER SUNDANCE SHALE
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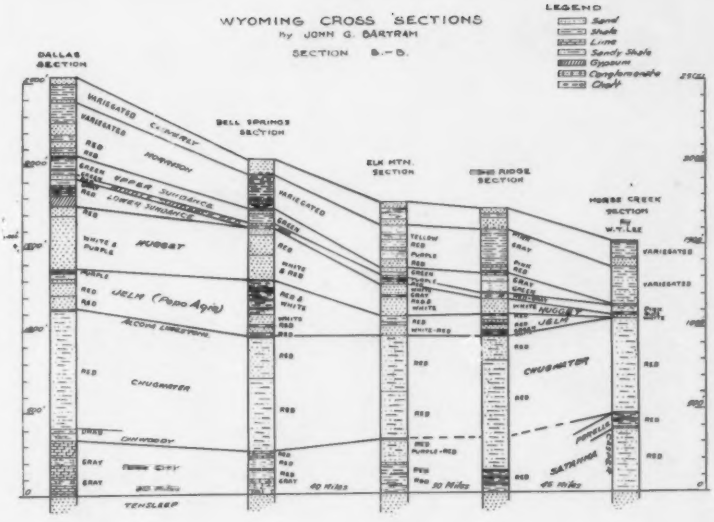
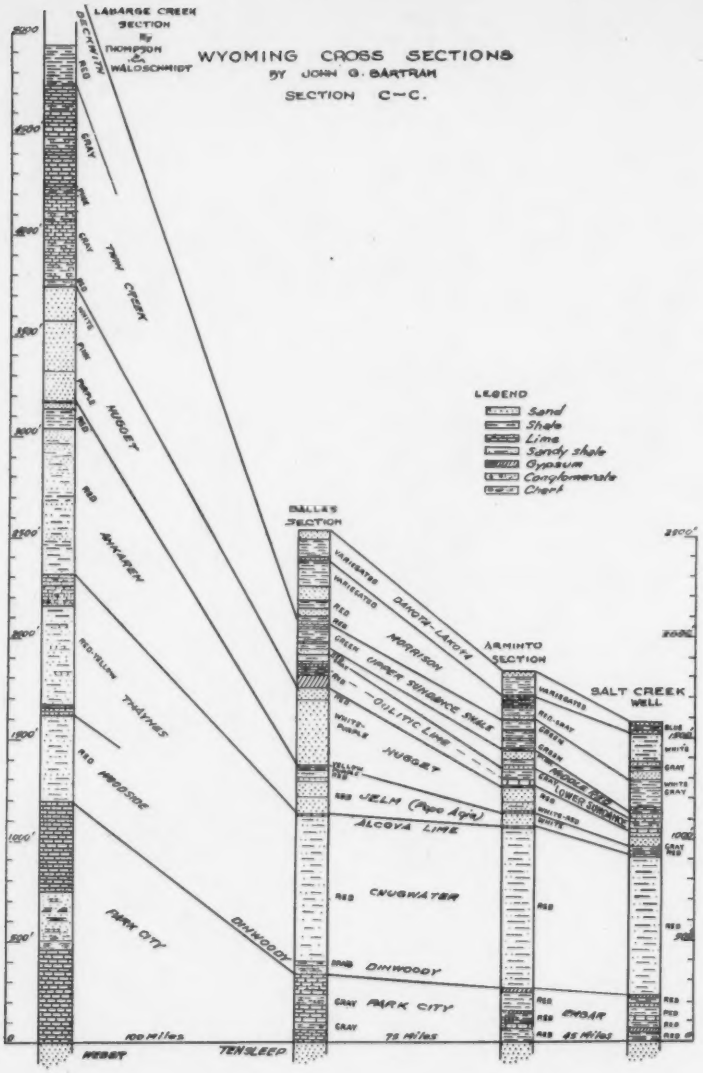

WYOMING CROSS SECTIONS
by JOHN G. BARTRAM
SECTION B.–B.
LEGEND
Sand
Shale
Lime
Sandy Shale
Gypsum
Conglomerate
Chert
DALLAS SECTION
BELL SPRINGS SECTION
ELK MTN. SECTION
CONE RIDGE SECTION
MORSE CREEK SECTION
by W.T. LEE
VARIEGATED
RED
GREEN
GRAY
WHITE & PURPLE
NUGGET
JELM (Pogo Agrio)
ALCOVA LIMESTONE
CHUGWATER
CHINKORDY
TENSLEEP
SATHANIA
PIPELLE
40 Miles
10 Miles
45 Miles


WYOMING CROSS SECTIONS
BY JOHN G. BARTRAM
SECTION C–C.
LABARGE CREEK SECTION By THOMPSON & WALDSCHMIDT
BELLE'S SECTION
ARMINTO SECTION
SALT CREEK WELL
LEGEND
Sand
Shale
Lime
Sandy shale
Gypsum
Conglomerate
Chert
TWIN CREEK
NUGGET
ANKAREH
THAYNES
WOODSIDE
PARK CITY
DINWOODY
GRAY CANYON
MORRISON
DANOW-LAKOTA
UPPER SUNDANCE SHALE
OOLITIC LIME
NUGGET
JELM (Pipe Agra)
ALCOVA LIME
CHUGWATER
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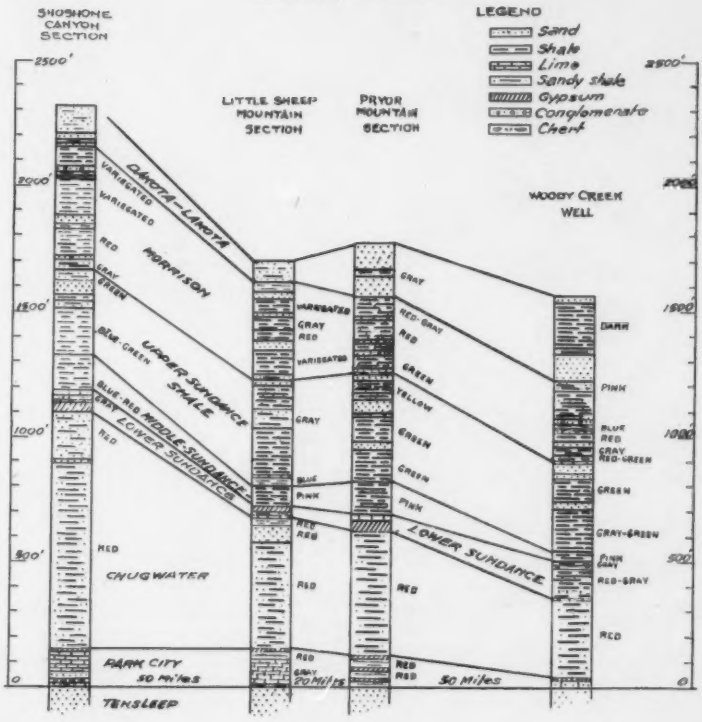
ENDAR
PHOSPHORIA BEDS
LOWER SUNDANCE
100 Miles
75 Miles
45 Miles
WEBER
TENSLEEP


WYOMING CROSS SECTIONS
By JOHN G. BARTRAM
SECTION D—D.
SHOSHONE CANYON SECTION
LEGEND
Sand
Shale
Lime
Sandy shale
Gypsum
Conglomerate
Chert
LITTLE SHEEP MOUNTAIN SECTION
PRYOR MOUNTAIN SECTION
WOODY CREEK WELL
2500'
2000'
1500'
1000'
500'
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DAKOTA-LAKOTA
MORRISON
UPPER SUNDANCE SHALE
MIDDLE SUNDANCE
LOWER SUNDANCE
CHUGWATER
PARK CITY
TENSLEEP
VARIEGATED
RED
GRAY
GREEN
BLUE-GREEN
BLUE-RED
GRAY
LOWER SUNDANCE
50 Miles
20 Miles
50 Miles
PINK
YELLOW
BLUE
RED-GRAY
GRAY-GREEN
DARK